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UNIVERSITY OF ALBERTA

THE QUATERNARY HISTORY OF THE CYPRESS HILLS AND ADJACENT AREAS IN ALBERTA AND SASKATCHEWAN.

BY

JOHN J. KULIG

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY.

DEPARTMENT OF GEOLOGY

Edmonton, Alberta
Spring 1995

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FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled The Quaternary Geology of the Cypress Hills and adjacent areas in Alberta and Saskatchewan submitted by John J. Kulig in partial fulfillment for the degree of Doctorate of Philosophy.

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November 25, 1994

This thesis is dedicated to my mother, my wife, and my Aunt and Uncle Johnny and Helen Nimcan.

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There were many times when I thought that this thesis would never be finished. It was definitely a trying experience. I see many avenues where continued additional research would answer many questions. In many ways that is what science is all about asking questions which when answered generate other questions. Now if I can only find some funding.......

Abstract

The objectives of this investigation into the surficial deposits covering the Cypress Hills and the adjacent plains in Saskatchewan are, (1) to determine the genesis of the diamicton units in the study area that were described in the field and analyzed in thin-section, (2) to determine the origin of the Frenchman channel, (3) to develop a deglaciation pattern that updates previously published regional chronologies and deglaciation sequences from adjacent areas. Two glacigenic units are present south of the Saskatchewan Cypress Hills along the border between Saskatchewan and Montana, a widespread thick (>10m) subglacial meltout till underlying a subaerial glacigenic debris-flow complex. The meltout till's widespread distribution reflects: (1) a highly concentrated basal debris layer, (2) the overlying subaerial debris-flow unit's role as an insulating mantle that slowed the rate of meltout, and (3) the maintenance of drainage away from the ice front precluding formation of large ice-contact glacial lakes in which glaciolacustrine depositional processes would have predominated. Glacigenic deposits in the Canal, Gilchrist, and Cypress Lake sections, on and near the southern flanks of the hills, consist of diamicton beds containing diamicton pebbles and irregular intraclasts of normally graded silts. The diamicton beds are intercalated with sand, silt, and clay laminae, which are deformed by dropstones. They were deposited in lakes impounded at the Late Wisconsinan Maximum when the ice (West, Gap, and East lobes) surrounding the Cypress Hills coalesced blocking the local drainage. Catastrophic release of these lakes during the initial stages of retreat from the maximum of the Late Wisconsin Glaciation, cut the Frenchman channel in its sidehill position along the Cypress Hills The outburst began with the release of glacial Lake Graburn near Merryflat, which triggered the subsequent outbursts of glacial lakes Cypress, Belanger, Robsart, and Blacktail.

Thin-section analysis of sediments from the study area and central Alberta revealed that the subglacial meltout till had well to strongly developed skelsepic and lattisepic plasma fabrics with clean grain boundaries. Glacially derived debris-flows, whether deposited subaerially or subaquatically, have insepic plasma fabrics and obscured grain boundaries.

The ice of the last part of the Wisconsin Glaciation was the most extensive ice. This may reflect formation of a western ice divide not present during earlier glaciations. Erratics scattered over the surface beyond the limits of this ice are not the remnants of older more extensive glaciations but material ice rafted across the lakes impounded at the maximum. Two advances (Middle Creek and Altawan) were newly recognized in the area south of the Cypress Hills. Also the maximum of limit of the Late Wisconsin Glaciation has been redefined and given a new name, the Underdahl Advance. During deglaciation the East Lobe was largely immobile, whereas the West Lobe underwent several extensive advances and retreats and is more similar to a continuously surging margin. The West Lobe was probably derived from a different ice divide from the East Lobe.

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Chapter 1: **Introduction**.

An understanding of the glaciation of the Cypress Hills region and the adjacent plains is crucial to reconstructing the glaciation sequence of southwest Alberta and Saskatchewan. McConnell (1885) first described the surficial deposits in the Cypress Hills region and later studies by Johnston et al. (1928), Bretz (1943), and Horberg (1956) added details. Westgate (1965, 1968) investigated the glacial geology of the Foremost map sheet that contains the Cypress Hills in Alberta. Whitaker (1965) completed a similar study of the Wood Mountain map sheet to the east. Klassen (1984, 1991, 1992) was the first to undertake detailed work in the area of the Cypress Hills in Saskatchewan. This study combines the results from the previous investigations with the results of detailed sedimentological analysis of the glacigenic sediment in the area to achieve a thorough understanding of the glacial geology and history of the Cypress Hills of Alberta and Saskatchewan and the adjacent terrain.

The five papers in this dissertation are based on section descriptions, air-photo analysis, and the distribution of such landforms as ice-marginal channels, spillways, outwash fans, and ice-thrust terrain. The objectives of the study were to (1) identify the varieties of glacigenic sediment present in the study area using both field investigation and thin-section analysis and develop depositional models to explain the origins of these sediments, (2) to compare glacial deposition in the study area to that of glacial deposits previously described in Alberta, (3) to determine the origin of the Frenchman channel and how this relates to the distribution of glacial deposits and ice lobes in the study area, (4) to use the sedimentological investigations and the Frenchman channel study to revise previously published Late Wisconsinan[1] glacial sequences for the region. The results of these objectives are presented and discussed in Chapters 2 through 6 and the overall conclusions summarized in Chapter 7.

---

[1] In his discussion of the Quaternary statigraphy of the Canadian Prairies, Fenton (1984) discussed the Early, Middle, and Late Wisconsinan substages and the reasoning behind the stratigraphy was detailed. Fulton et al. (1984) summarize the Quaternary stratigraphy of Canada and supply the following time ranges for the Early, Middle and Late Wisconsinan substages. Early Wisconsinan 75,000 to 64,000 BP, Middle Wisconsinan 64,000 to 25,000 BP, and Late Wisconsinan 25,000 to about 10,000 BP. Despite problems with correlation the stratigraphy presented by Fenton (1984) and Fulton et al. (1984) provides an excellent framework for further investigation and discussion.

Paper 1 (Chapter 2) details investigations into the sedimentological processes that deposited the glacigenic sedimentary units that cover most of the area near the border of Saskatchewan and Montana, between the Old Man On His Back and Boundary plateaus on the east, and Lodgepole Creek to the west (Fig. 2.1, 2.2). Two widespread units were observed in river cut exposures: a lower, subglacial meltout till unit, several tens of metres thick and an overlying subaerial glacigenic debris-flow complex of variable thickness. The preservation of a thick, widespread meltout till reflects three factors: (1) the presence of a highly concentrated debris layer near the base of the ice. This concentration of debris in this zone decreased the volume of meltwater released per unit of debris-laden ice melted and reduced the degree of deformation during meltout; (2) the insulating effect of the overlying subaerial, glacigenic debris-flow complex, that slowed the rate of melting, increased the likelihood of preserving the subglacial till of Unit 1, (3) the maintenance of regional drainage away from the ice ensured drainage of the deposit and prevented formation of ice-marginal lakes in which glaciolacustrine depositional processes (see Evenson et al. 1977; Gibbard 1980; Kulig 1985) would predominate.

Paper 2 (Chapter 3) contains descriptions of the Canal, Gilchrist, and Cypress Lake sections, on and near the south flanks of the Saskatchewan Cypress Hills. The sections contain diamicton intercalated with continuous sand and silt laminae. Sedimentary features such as diamicton intraclasts, rhythmite intraclasts, continuous sub-horizontal silt and clay laminae, and continuous, thin sub-horizontal diamicton layers, and silt laminae deformed beneath large phenoclasts (dropstones) demonstrate that the units are glaciolacustrine in origin. The glacial lakes in which the sediment was deposited formed when the three ice lobes (West, Gap, and East lobes) that surrounded the Cypress Hills during the Late Wisconsinan maximum coalesced, blocking regional drainage to the southeast (Fig. 3.1B). Ice rafting across these lakes explains the distribution of unweathered erratics beyond the limit of the last advance. Impounding these ice-contact lakes on and around the south slopes of the Cypress Hills required more extensive ice during the last part of the Wisconsin Glaciation than presented in Christiansen (1979), Clayton and Moran (1982), or Dyke and Prest (1987a). These lakes are also of critical importance to the formation of the Frenchman channel described in Paper 3.

Paper 3 (Chapter 4) presents evidence supporting the origin of the Frenchman channel by the catastrophic release of lakes impounded on or around the south flanks of the Cypress Hills (Fig. 4.5). The impounding of ice-contact lakes described in Paper 2 requires that ice of the East Lobe extend farther to the west and cover the area occupied by the channel. In previous interpretations the ice of the East Lobe terminated north of the channel. The Frenchman channel was therefore an interlobate feature incised gradually by meltwater draining across and along the southern slopes of the Cypress Hills (Christiansen 1979; Clayton and Moran 1982; Christiansen and Sauer 1988). The proposed new ice configuration indicates that the channel was not an interlobate feature. The characteristics of the channel match those of spillways formed by the catastrophic release of ice-dammed lakes (Kehew and Lord 1986, 1987). Since the head of the channel is at Merryflat, west of the lakes discussed in Paper 2, the catastrophic release of these lakes did not initiate the channel. Instead, formation of the channel began with the outburst of glacial Lake Graburn, located on the south flanks of the West Block of the Cypress Hills. This outburst triggered the release of glacial lakes Belanger, Cypress, and Robsart. Glacial Lake Robsart formed in the confluence area between the East and West lobes and drained through Palisades Coulee.

The spillway hypothesis contrasts with the subglacial megaflood theory in which the Frenchman channel is one of the final segments in the subglacial channel system that transported megafloods from the Northwest Territories to the Mississippi River (Shaw, pers. comm., 1994). The local subglacial and ice-proximal drainage of large ice dammed lakes suggests an alternative explanation for the formation of the subglacial fluvial features described in Shaw and Kvill (1984) and Shaw et al. (1989), and the scabland terrain described in Rains et al. (1993). Features north of the Cypress Hills and on the Swift Current plateau as evidence of a regional subglacial drainage event (Shaw 1994, pers. comm.), could also be formed by subglacial catastrophic drainage of the lake that was impounded immediately north of the Cypress Hills. Since the wastage of the ice, from Alberta and southern Saskatchewan, occurred in a downslope direction blocking regional drainage, ice-contact glacial lakes were commonly impounded along the ice margin. The catastrophic release of ice-marginal lakes at successively lower elevation may have generated the large scabland that extends from near the Cypress Hills to central Alberta. A combination of subglacial and subaerial outbursts from these lakes could have formed scablands that contain subglacial elements and spillway features. Overlapping of several successive catastrophic outbursts would form an apparently continuous scoured surface. This landscape is called a shingled scabland.

Paper 4 (Chapter 5) is a departure from detailed section work and regional analysis. It documents the application of thin-section analysis to determining the genesis of glacigenic diamicton units. The analysis indicates that glacigenic diamicton deposits formed by resedimentation (sub-aerial and sub-aquatic glacially derived debris-flows) lack well-developed matrix fabrics, contain few phenoclasts greater than fine pebble size in their matrix, and contained many rounded and subrounded diamicton intraclasts. They may also possess pebble-cored mud intraclasts, and several varieties of silt and rhythmite intraclasts formed during remobilization of previously deposited glaciolacustrine rhythmites. Thin-section examinations of the contacts between laminae and the over- and underlying diamicton layers allow the laminae to be distinguished from similar appearing linear features such as shear planes. The subglacial meltout till examined possesses less matrix, has moderate to well-developed skelsepic and skel-lattisepic plasma fabrics, and lacks the soft sediment intraclasts observed in the resedimented diamicton. The plasma fabrics observed in the subglacial meltout till may reflect compression beneath an overlying debris-laden ice mass.

Paper 5 (Chapter 6) presents a revised glacial sequence for the Cypress Hills of Alberta and Saskatchewan. It is based on the fieldwork contained in the first four papers and additional sections investigated during the study. The new sequence modifies the glacial sequence developed by Westgate (1965, 1968) for the Alberta part of the Cypress Hills and extends it into the adjacent area of Saskatchewan. Chapter 6 also revises the broad regional sequences published by Clayton and Moran (1982), Fullerton and Colton (1986), and Dyke and Prest (1987a, b). The Elkwater drift (Westgate 1965, 1968, 1972) and "residual drift landscape complex" (Klassen 1992), previously believed to be the remnants of an older more extensive glacial advance, are reinterpreted as ice-rafted deposits. Two newly named advances, the Middle Creek and Altawan, are identified. Additionally since the limits of the Green Lake Advance (Westgate 1972) are redefined, the advance at the Late Wisconsinan maximum is renamed the Underdahl Advance.

The timing of the deglaciation sequence closely follows the chronology established by Clayton and Moran (1982), that was based on radiocarbon dates only from wood and excludes dates obtained from disseminated organic matter in which incorporation of old carbon leads to anomalously old dates (Teller 1987; MacDonald et al. 1989). Only deposits from the Late Wisconsinan ice have been reliably identified in the Cypress Hills area. The absence of evidence for earlier ice advances supports (but does not prove) the hypothesis of Liverman et al. (1989) and Young et al. (1993) that the Late Wisconsinan ice was the most extensive ice sheet on the Canadian Prairies. It is possible that a western ice divide formed during the Late Wisconsinan Glaciation and not during previous glaciations. The movement of the ice and its distribution over Alberta and Saskatchewan during the Late Wisconsinan was therefore very different from the flow patterns and distributions of earlier glaciations.

A short note here on coordinates for the sections and sites visited during fieldwork. Two systems are supplied, the common township, range section, quarter section system and the military grid system. The township-range designations provide immediate basic locations so that people familiar with this system can place the site approximately. To give more accurate locations the 16 quarters in a section have been in turn divided into 4. This allows placement of a section to a 400 metre square (1320 ft X 1320 ft). For some sections and sites this is still coarse so the military grid designation was also determined. This system starts with the map series, followed by a grid zone designation. The 100,000 metre square identification is next. Finally the easting and northing numbers are provided. This number is written as one continuous string. Thus Battle Creek 1 (BC1) is 72F/3,12UXK153287: 72F/3 the Lyons Creek 1 to 50,000 map sheet; 12U is its grid zone, XK is its reference square, 153 is the easting and 287 is the northing. This allows a point to be determined to within 100 metres.

References cited.

Bretz, J.H., 1943, Keewatin end moraines in Alberta, Canada: Geological Society of America Bulletin, v. 54, p. 31-52.

Christiansen, E.A., 1979, The Wisconsinan glaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences: v. 16, p. 913-938.

Christiansen, E.A. and Sauer, E.K., 1988, Age of the Frenchman Valley and associated drift south of the Cypress Hills, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 25, p. 1703-1708.

Clayton, L. and Moran, S.R. 1982, Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55-82.

Dyke, A.S., and Prest, V. K., 1987a, Late Wisconsinan and Holocene history of the Laurentide ice sheet: Geographie physique et Quaternaire, v. 41, p. 237-263,

Dyke, A.S., and Prest, V. K., 1987b, The Late Wisconsinan Maximum, Geological Survey of Canada map 1702a, scale 1:5,000,000.

Evenson, E.D., Dreimanis, A., Newsome, J.W., 1977, Subaquatic flow till: a new interpretation for the genesis of some laminated till deposits: Boreas, v. 6, p. 115-133.

Fenton, M.M., 1984, Quaternary stratigraphy of the Canadian Prairies, *in* Fulton, R.J. ed., Quaternary Stratigraphy of Canada - a Canadian Contribution to IGCP Project 24. Geological Survey of Canada, Paper 84-10. p. 57-68.

Fullerton, D.S., and Colton, R.B., 1986, Stratigraphy and correlation of the glacial deposits on the Montana Plains. *in* Richmond, G.M. and Fullerton, D.S., eds., Quaternary Glaciations of the United States: Quaternary Science Reviews, v. 5, p. 69-82

Fulton, R.J., Fenton, M.M., and Rutter, N.W., 1984, Summary of Quaternary Stratigraphy and History, western Canada, *in* Fulton, R.J. ed., Quaternary Stratigraphy of Canada - a Canadian Contribution to IGCP Project 24. Geological Survey of Canada, Paper 84-10. p. 70-83.

Gibbard, P., 1980, The origin of the Catfish Creek Till by basal melting: Boreas, v. 9, p. 71-80.

Horberg, L., 1952, Pleistocene drift sheets in the Lethbridge region, Alberta, Canada: Journal of Geology, v. 60, p. 303-330.

Johnston, W.A. and Wickenden, R.T., 1931, Moraines and glacial lakes in southern Saskatchewan and southern Alberta, Canada: Transactions of the Royal Society of Canada, Section 4: Geological Sciences including Mineralogy, v. 5, p. 29-44.

Kehew, A.E., and Lord, M.L., 1986, Origin and large-scale erosional features of glacial-lake spillways in the Northern Great Plains: Geological Society of America Bulletin, v. 97, p. 162-177.

Kehew, A.E., and Lord, M.L., 1987, Glacial-lake outbursts along the mid-continent margins of the Laurentide ice-sheet. *in*: Mayer, L., and Nash, D., eds., Catastrophic Flooding, Allen and Unwin, Boston, p. 95-120.

Klassen, R.W., 1984, Quaternary geology of southwestern Saskatchewan: *in* Current Research, Part A, Geological Survey of Canada. Paper 84-1A, p. 641-642.

Klassen, R.W., 1991, Surficial geology and drift thickness, Cypress Lake (72F) Saskatchewan: Geological Survey of Canada, Map 1766A (1:250,000).

Klassen, R.W., 1992, Nature, origin, and age relationships of landscape complexes in southwestern Saskatchewan: Geographie physique et Quaternaire, v. 46, p. 361-368.

Kulig, J.J., 1985, A sedimentation model for the deposition of glacigenic deposits in central Alberta. [M.Sc. thesis], Edmonton, Alberta, Canada, University of Alberta, 245 p.

Liverman, D., Catto, N.R., Rutter, N.W., 1989, Laurentide glaciation in west central Alberta: a single (Late Wisconsinan) event: Canadian Journal of Earth Sciences, v. 26, p. 266-274.

MacDonald, G.M., Beukens, R.P. , Kieser, W.E., and Vitt, D.H., 1987, Comparative radiocarbon dating of terrestrial plant and macrofossils and aquatic moss from the 'Ice Free Corridor' of western Canada: Geology, v. 15, p. 837-840.

McConnell, R., 1885, Report on the Cypress Hills, Wood Mountain, and adjacent country: Geological Survey of Canada, Annual Report, 1C.

Rains, B., Shaw, J., Skoye, R., Sjogren, D., and Kvill, D., 1993, Late Wisconsinan mega-flood paths in Alberta: Geology, v. 21, p. 323-326.

St.Onge, D.A., 1972, Sequence of glacial lakes in north-central Alberta: Geological Survey of Canada Bulletin 213, 16 pp.

Shaw, J., and Kvill, D., 1984, A glaciofluvial origin for drumlins of the Livingstone Lake area, northern Saskatchewan: Canadian Journal of Earth Sciences, v. 21, p. 1442-1459.

Shaw, J., Kvill, D., and Rains, B., 1989, Drumlins and catastrophic subglacial floods: Sedimentary Geology, v. 62, p. 177-202.

Teller, J.T., 1989, Importance of the Rossendale site in establishing a deglacial chronology along the southwestern margin of the Laurentide Ice Sheet: Quaternary Research, v. 32, p. 12-23.

Westgate, J.A., 1965a, The surficial geology of the Cypress Hills area, Alberta: Alberta Research Council Preliminary Report 65-2, 6pp.

Westgate, J.A., 1968, Surficial geology of the Foremost - Cypress Hills area, Alberta: Research Council of Alberta Bulletin 22. 122 pp.

Westgate, J.A., 1972, The Cypress Hills. *in* Quaternary geology and Geomorphology between Winnipeg and the Rocky Mountains. International Geological Congress, 24th Session Montreal Quebec, Guidebook for Excursion C-22 p. 50-62,

Whitaker, S.H. 1965, Geology of the Wood Mountain Area (72G) Saskatchewan. [Ph.D. Thesis], Urbana, IL, University of Illinois, 240p.

Young, R. R., Burns, J.A., Rains, R.B., Smith, D.G., 1993, The Late Wisconsinan Laurentide Ice sheet and its relationship to a theoretical "ice-free" corridor: *in* Program with abstracts, Geological Association of Canada, Joint Annual Meeting May 17-19, 1993, Edmonton, Alberta, p. A-113.

Chapter 2: The genesis of glacial sedimentary units in southwest Saskatchewan.

A version of this paper will be submitted for publication in Boreas.

Introduction

This sedimentological study is the first to examine in detail the glacial sediment exposed between and long Battle and Lyons creeks (Fig. 2.1, 2.2) of southwest Saskatchewan. The objectives of the investigation were to determine the genesis of the two glacigenic sedimentary units observed in exposures in river cuts, creeks, and isolated exposures throughout the field area and then place the units into the local and regional glacial contexts. The sedimentological analysis shows that the lower unit is a thick subglacial meltout till (up to 35 m) that underlies a subaerial glacigenic debris-flow complex. The factors controlling the widespread distribution of this meltout till, are examined. The possibility that two broad glacial depositional settings are present on the Prairies is also discussed.

Geologic background and fieldwork

The study area is south of the Cypress Hills in the southwest corner of Saskatchewan (Fig. 2.1). The glacial sediment was deposited from an ice lobe that flowed southeasterly around the Cypress Hills of Alberta and then spread eastward and southward (Westgate 1968; Klassen 1992). During deglaciation, the upper part of the surficial sediment was locally reworked by meltwater, forming a flat to gently irregular eroded till plain with local relief of 1 to 5 m (Klassen 1991, 1992). No collapsed hummocks are present in the till plain in the study area. A single spillway north of the area, the Frenchman channel, directed meltwater eastwards, away from the study area (Fig. 2.1).

The sections studied for this paper are river cut exposures that range from 10 to 45 m high along the Lyons and Battle creeks near the Montana and Saskatchewan border (Fig. 2.1). A total of 5 sections along the Battle Creek and 4 along the Lyons Creek were investigated (Fig. 2.1, 2.2). The vertical exposures formed when these creeks cut into the flat to gently undulating surface of the till plain. The tops of all the sections are flat, with a maximum surface relief of 1-2 m. BC1: West of 3, T1, R26, S4, 1/4 sect., South half of 3 to SW quad. of 2, BC2: T1, R26, S4, 1/4 sect., NE quadrant of 4 to SW quadrant of 6, BC3: T1, R26, S4, 1/4 sect., SW quadrant of 1, BC4: T1, R26, S21, 1/4 sect., SE quadrant of 2, BC5: T1, R26, S33, 1/4 sect., SW quadrant of 13 to NE quadrant of 13. The UTM grid coordinates for the sections are Battle Creek 1 (BC1) are 72F/3,12UXK153287, BC2: 72F/3,12UXK149286 to 151291, BC3: 72F/3,12UXK159286, BC4: 72F/3,12UXK155333 to 156334, BC5: 72F/3,12UXK145377 to 147379. Investigation was primarily undertaken on BC1, BC2 and BC3 and supplemented with observations on BC3, BC4 and BC5. The LSD designations of the Lyons Creek sections are: LC1: West of 3, T1, R25, S14, 1/4 sect., NW quadrant of 14, LC2: T1, R25, S23, 1/4 sect., SW quadrant of 3, LC3: T1, R25, S23, 1/4 sect., NE quadrant of 4, LC4: T1, R25, S23, 1/4 sect., SE and NE quadrants of 11.Lyons Creek sections 72F/3,12UXK282337 (LC1), 72F/3,12UXK282338 (LC2), 72F/3,12UXK278337 (LC3), 72F/3,12UXK282342 (LC4).

Fieldwork consisted of section description and photography. The sections were sampled for bulk texture, and block samples were removed for the production of thin sections. Pebble fabrics were measured using clasts with a-axis longer than 1.0 cm and with a/b axial ratios larger than 1.5. These were plotted as bi-directional rose diagrams to determine their degree of multimodality and interpreted following the methods described in Mark (1973, 1974).

Previous work

Prior to investigations by Klassen (1984, 1991, 1992) and Klassen and Vreeken (1985, 1987), the Quaternary glacial history of the Cypress Hills in southwest Saskatchewan and the plains to the south was reconstructed from regional studies and early surficial deposit descriptions by McConnell, (1885), Johnston and Wickenden (1931), Bretz (1943), and Johnson et al. (1948). Westgate (1965a, b, 1968) described the deposits and glacial stratigraphy of the adjacent Foremost map sheet in southeast Alberta and identified five drift sheets (Elkwater, Wild Horse, Pakowki, Etzikom, Walsh). Only the Elkwater drift, a thin patchy till on the flanks of the Cypress Hills, was interpreted to be older than Late Wisconsinan (Westgate 1968, 1972). The Wild Horse drift (Westgate 1972) deposited during the Late Wisconsinan maximum must have extended into the study area, but the remaining drift sheets were confined to Alberta. East of the study area, Whitaker (1965, 1976) interpreted the surficial deposits of the adjoining Wood Mountain map sheet to be Late Wisconsinan. Both Westgate (1968, 1972) and Whitaker (1965) placed the Late Wisconsinan ice margin south of the Cypress Hills. In contrast, Stalker (1977) and Stalker and Harrison (1977) concluded that the southern limit of the Late Wisconsinan ice was the Lethbridge moraine, which they correlated to the Pakowki moraine west and north of the Cypress Hills. The surficial deposits south of the Cypress Hills were therefore older than Late Wisconsinan. Regional stratigraphic considerations led Christiansen (1979), Clayton and Moran (1982), and Fullerton and Colton (1986) to dispute Stalker's Late Wisconsinan margin and place it to the south, in Montana. From soil, tephra, and loess investigations in the Elkwater, Alberta area (Fig. 2.1), Vreeken (1986) concluded that the surface deposits surrounding the Alberta Cypress Hills were Late Wisconsinan. Analysis of the nature, thickness, and radiocarbon dates in nonglacigenic surface units led Klassen and Vreeken (1987) to also conclude that the surface south of Cypress Hills was Late Wisconsinan. Christiansen and Sauer (1989) used radiocarbon dates from wood contained in alluvial sediments within the Frenchman channel to interpret it as a Late Wisconsinan feature. Klassen (1992) placed the study area into his "last advance landscape complex". The deposits in the study area would therefore be the temporal equivalents of the Late Wisconsinan Battleford Formation (Christiansen, 1968, 1992).

Description of the sedimentary units

Two sedimentary units were recognized in the sections: a lower thick unbedded diamicton unit and an upper discontinuous unit composed of more than 50% sorted sediment interlayered with diamicton beds. In the following description, observations made at BC1 (Fig. 2.2), the most extensive, complex, and complete section are supplemented by observations made at the smaller Lyons Creek sections (Fig. 2.1, 2.2).

Unit 1. Diamicton with bedrock blocks, slabs and isolated sand lenses

Unit 1 is 25 to 35 m thick and composed predominantly of dark gray, clay-rich diamicton. Sand laminae are rare, but isolated sand lenses, sand and gravel pods are abundant and found throughout the section. Bedrock slabs and blocks of variable size are distributed throughout the unit.

The unit extends to the surface or is overlain by Unit 2. Where Unit 1 extends to the surface, it is either unbedded with a platy breaking pattern or crudely stratified with a crumbly breaking pattern in which individual shards are pebble-sized. The unit has a recessive gullied appearance (Fig. 2.5).

Diamicton

Unbedded diamicton, uniform in grain size and color (dark gray 4.5Y4/0 moist) makes up about 85% of the unit (Fig. 2.3, 2.5, 2.6, 2.7). The average texture of the diamicton, minus the pebble and coarser fraction, is 41% clay, 35% silt, and 25% sand. The diamicton has poorly defined subhorizontal partings and a tabular to platy breaking pattern. Phenoclasts pebble-sized or larger, total 8 to 12% of the diamicton volume and are predominantly granite, quartzite, and shale, with a minor carbonate component. Throughout the sections are flat-iron shaped, weakly striated, shale phenoclasts, with a axes of 1 to 3 cm. The pebbles make up 1 to 2% of the total diamicton volume, but localized concentrations of shale phenoclasts up to 30% were observed.

The diamicton has a dense, clay-dominated matrix. Moderate to well-developed orientations of the clays around lithoclasts and sand grains (skelseptic fabric, Brewer 1976) were observed in thin-section. No soft-sediment intraclasts (diamicton or rhythmite intraclasts), nor pebble-cored intraclasts were observed in thin-section or in outcrop. In thin-section, the shale pebbles had uniform extinction angles and showed no indication of shearing or penetrative deformation.

Seventeen fabrics were measured in this unit (Table 2.1). A plot of S_1 eigenvalues versus trend (Fig. 2.4) indicates a clustering around 270^o. A grand mean of all the fabric measurements produced a trend of 285.8^o and a plunge of 19^o, with an S_1 eigenvalue of 0.69.

Bedrock components: stringers, slabs and blocks

Bedrock components are important in Unit 1 (Fig. 2.7, 2.8, 2.9, 2.10). They vary from thin bedrock stringers (Fig. 2.7), a few millimetres to centimetres in width and tens of centimeters in length; to tabular slabs, 10 to 30 cm thick and up to 15 m in length (Fig. 2.8); to blocks up to 7 m long, 1.5 m thick and more than 2m in thickness (Fig. 2.7, 2.8, 2.9). Bedrock blocks and slabs extend over 1 m into the section.

Stringers

Stringers (Fig. 2.7) are 1 to 20 mm thick and commonly less than 20 cm long but some can be traced for several metres. They are usually gray-green silty sandstone or orange iron-stained sandstone and have sharp upper and lower contacts. Occasionally flame-like structures extend a few centimeters from the stringers into the surrounding diamicton. One bedrock stringer observed in thin-section had sharp contacts and no shearing, streaking, or diamicton incorporation into the stringer. The trend and plunge of the stringers match that of the bedrock slabs.

Slabs

Slabs of shale or gray-green or orange sandstone, are thicker and longer than stringers (Fig. 2.8). Slabs 8 to 10 m long and 10 to 15 cm thick may thin to 5 to 10 mm in thickness. Shale slabs commonly have bright orange iron-staining along joint faces and a puffed botryoidal surface texture caused by clay expansion. The slight color difference between the diamicton and shale slabs makes their recognition difficult. Some thin shale slabs are contorted. Five thin-sections containing the contact between a shale or sandstone slab and diamicton showed intrusion of phenoclast-poor diamicton into the slab. The thin-sections often showed distinct rounded and oval sub-units with uniform extinction angles within the shale slabs.

Section BC1 has the highest concentration of yellow and orange sandstone slabs (Fig.2.3, 2.8). The slabs are continuous for 5 to 15 m, make sharp contacts with the enclosing diamicton, and are readily identified by their distinct color. Several smaller slabs can be linked, to reconstruct a single bedrock slab that extends across the section from its center to near its top (Fig. 2.8). Three such extensive slabs were traced through the section. The slabs have a consistent west to southwesterly long-axis trends and plunges of between 4 and 6 degrees (Fig.2.3, 2.8). Locally, thin diamicton partings split the slabs into sub-parallel units. Small (10 to 15 cm long), flame-like protrusions of bedrock into the diamicton are observed, but are not common. In several locations, the slabs are broken by high-angle normal faults with 10 to 15 cm offsets.

Blocks

There are numerous large bedrock blocks in all the sections. In places, these large blocks form the bulk of the unit. They range in size from 1 to 2 m^3, to blocks, 5 to 6 m wide and 4 m long that extend an unknown distance into the section. Blocks (2 to 3 m high and 2 to 3 m wide) are commonly observed at the top of all sections. The blocks are generally composed of shale but light yellow sandstone blocks are also seen. One highly jointed soft-shale block in BC1, 1.5 to 1.7 m thick, more than 2 m deep, and about 10 m in length possesses an extensive joint network highlighted by iron staining, that permeates the outer margins of the block. Thin section examination of diamicton stringers in the outer 40 cm of the shale block indicates that granular material coarser than fine sand is not present in the stringers (Fig. 2.10). Although the block appears intact, well-rounded quartzite pebbles derived from local Miocene gravels are encased within it (Fig. 2.8). Several other smaller shale blocks in BC1, LC1, and LC2 contain thin diamicton stringers along their margins and isolated quartzite phenoclasts within themselves. Neither diamicton pods nor stringers were associated with these isolated phenoclasts.

Sand and gravel lenses, and sand laminae

Lenses of sand and gravel, distributed throughout Unit 1, range from 10 to 250 cm thick and 2.5 to 15 m in length. Numerous thin tabular sand lenses (10 to 15 cm and 2 to 3 m long) are present throughout LC1 and LC2 are rare in Unit 1 of BC1. The lenses commonly contain planar and cross-stratified ripples of fine to medium-grained sand but planar cross beds are also present. The lenses have sharp conformable upper and lower contacts with the surrounding diamicton.

A large (2.5 m by 1.5 m) isolated sand lens in LC3 contains sedimentary structures that are contorted and displaced downwards tens of centimeters along its nearly vertical lateral contacts (Fig. 2.11). Sedimentary structures along the sides of other lenses are also deformed upwards. Thin contorted diamicton diapirs often extend from the surrounding diamicton into the sorted sediment.

On the east flank of BC1, a sand pod (2.5 m at its thickest and 10 m long), composed mostly of fine and medium sand, contains a 50 cm thick planar cross-bedded gravel layer formed of well-rounded shale pebbles. Along the left margin, the pod terminates abruptly in diamicton. Strata near the edge of the sand pod are contorted and downwarped. Westward the pod thins to 20 cm (over 8 m distance) and is overlain by a 30-cm-thick diamicton layer that separates the sand pod from an orange iron-stained sandstone slab. The trend and plunge of this slab are unaffected by the proximity of the sand lens and match those of the other slabs in Unit 1.

Sand lenses and bedrock slabs are also closely associated in the center of BC1, where a tabular sand lens 15 cm thick and 2 m long was observed 15 cm below a 10 m long gray-green sandstone slab. The lens contains planar sand laminae and possesses sharp upper and lower contacts with the surrounding diamicton. Both the slab and sand lens are undeformed.

Quartzite boulders and cobbles (10 cm diameter) protrude from diamicton into underlying fine grained sand lenses (Fig. 2.12). Beneath one of these phenoclasts was a shallow scour (less than 3 cm deep) incised into the sand lens containing granules and diamicton intraclasts. Where another phenoclast extended from the diamicton through the underlying sand lens and into the underlying diamicton, a shallow scour, containing small pebbles and diamicton intraclasts, was observed in the underlying diamicton layer.

Genesis of Unit 1

Although diamicton is the largest component of Unit 1, evidence from the sand lenses and bedrock components facilitates genetic interpretation of the unit. Isolated sand lenses with intact primary bedding structures are observed throughout Unit 1. The unit therefore is unlikely to have formed by lodgement as the plastering-on associated with lodgement would have sheared and deformed the bedding structures within the lenses. Furthermore, absence of shear features in the bedrock blocks, slabs and shale pebbles makes it unlikely that they were thrust into place. Marcussen (1975) reported the streaking and smudging of soft lithologies in units interpreted as lodgement tills. In thin-section also, there is no evidence of shearing and shale pebbles and soft sandstone pebble-sized clasts are undeformed. A passive method of deposition is therefore indicated.

Isolated sand and gravel lenses with intact primary sedimentary structures interbedded within diamicton are common in till interpreted to be of meltout origin (Boulton 1970, 1971, 1972; Haldorsen and Shaw 1982; Shaw 1987). They are also common in glacial debris-flow complexes (Hartshorn 1956; Boulton 1972; Lawson 1979). The thickness of the unit and low volume of sorted sediments are inconsistent with formation by overlapping glacially-derived debris-flows. Dragfolding of bedding planes at the margins of large sand lenses in BC1 and LC2 and LC3 is consistent with unequal settling of the lenses as underlying buried ice melted. A debris-flow origin could not preserve the 10 to 15 m long, thin, bedrock slabs that extend from the center of Unit 1 to its top, and have consistent trends and plunges. The continuous nature of the bedrock slabs in BC1, indicate that there are no hidden sedimentary breaks in Unit 1 there and therefore a single process was responsible for deposition of Unit 1. Stacked slabs are commonly considered to reflect glaciotectonics (Banham 1977) or various freeze-on mechanisms (Weertman 1963; Moran 1971; Clayton and Moran 1974; Bluemle and Clayton 1984). These are entrainment mechanisms and may have little relationship to the mechanism of final deposition (Aber 1985; Rusczyñska-Szenajach, 1987, 1988). No shear planes or thrust features were observed in section. Thin-sections from diamicton-bedrock slab contacts also show no trace of shear bands or streaking. Ice thrusting therefore is unlikely to have deposited the bedrock slabs. Since all the exposures are beneath a till plain without collapsed hummocks and other features expected with large volumes of debris-flow sedimentation. This further decreases the likelihood that Unit 1 is composed of overlapping debris-flows.

The evidence is more in keeping with a meltout origin for the unit. The boulder scours located beneath cobbles and boulders too large to have been transported by the meltwater that deposited the sand surrounding these cobbles and boulders, are said to be characteristic of meltout tills (Shaw 1982, 1983, 1987). Scours form when meltwater flowing beneath ice is forced to flow past boulders projecting from the ice roof. The boulders cause flow perturbations that erode the underlying sediment. Upon meltout, the boulders are lowered into the scour (Shaw 1982, 1983). Scours beneath phenoclasts partially encased in diamicton are Shaw's (1983) Style 2 scours, whereas, those extending through the sand lens into the underlying diamicton and containing a lag of diamicton pebble intraclasts, are Shaw's (1983) Style 4 scours.

The sand lenses, boulder scours, and dragfolds of bedding planes all point to a meltout origin for Unit 1. The normal faults extending from the diamicton through the sand lens in LC3 and BC1 are similar to those believed to have been generated by the meltout of buried ice (Rust and Romanelli 1975; Stone 1976; Shaw 1982). Sand lenses in close association with bedrock slabs further support a meltout origin. The stacked slabs could not have been thrust into position by an advancing ice sheet (Bluemle and Clayton 1984) as this would have disrupted the ripple and planar crossbeds in the sand lenses only 10 to 30 cm beneath the bedrock slabs.

Unit 1 could not have a debris flow origin because stacked sediment flows would not preserve the soft bedrock slabs or their orientation. The bedrock slabs and the primary bedding features of the associated sand lenses could not have survived lodgement. Only passive subglacial meltout could have emplaced the bedrock slabs in direct association with meltwater deposits with intact primary structures and sand lenses containing boulder scours. The slabs were therefore entrained elsewhere, transported to the deposition area, and deposited by passive basal meltout. Identifying the entrainment process (thrusting Bluemle and Clayton 1984; Aber 1985; or freezing on: Weertman 1963; Moran 1971; Clayton and Moran 1974) is not crucial to determining the process of deposition. As basal transport of only a few tens of kilometers would easily destroy the thin (10 to 20 cm), soft, deformable sand and shale slabs and blocks (Aber 1985; Broster and Seaman 1991), the bedrock slabs in the sections must be locally derived. The Cretaceous Bearpaw Formation that outcrops throughout the area (Furnival 1948; Whitaker 1976) is the most likely source. Compressive flow conditions generated as the eastward-moving ice lobe flowed up the western flank of the Old Man On His Back Plateau (Fig. 2.1) may have stacked slabs within the ice.

Rusczyñska-Szenajach (1987) described similar bedrock slabs and blocks detached by freeze-on processes, transported within the ice and then deposited by meltout. They had sharp edges and intact internal features and are similar to the slabs in BC1. Elson (1961) and Lawson (1979a, 1989) stated that distinctive, tabular debris layers in the ice are usually disrupted and deformed during meltout due to the irregular distribution of debris. Because the thin, deformable bedrock slabs are preserved with only minor offsets, have a small degree of disruption, and have a common trend and plunge, the debris in the basal ice must have been concentrated and uniformly distributed. The thickness of the unit (up to 35 m) also suggests this. A large debris concentration with a regular distribution may indicate entrainment of diamicton slabs as well as bedrock slabs into the ice. Gillberg (1977) suggested that till is a complex mix of glacially eroded rock material and other sediment types that have undergone several cycles of deposition and reincorporation by an over-riding ice sheet. Elson (1989) described a deformation till in which previously deposited material underwent varying degrees of reworking by the over-riding ice. Ronnert (1992) similarly concluded that diamicton slabs were incorporated into the debris layer of the ice and then melted out. It is probable that the sections in the study area contain numerous diamicton slabs. Block inclusions of diamicton would be difficult to identify if there were no distinct layers separating them (Lawson 1979a; Aber 1985; Ronnert 1992).

The bedrock slabs generally show very little deformation, possess a common trend and plunge, and are associated with undeformed sand and gravel layers. To preserve thin bedrock slabs a concentrated debris layer is needed. Re-incorporation of slabs of previously deposited debris could supply the volume of englacial sediment. Entrainment, transport, and deposition 'en masse' of diamicton slabs could also explain the lack of consistent strong phenoclast fabrics that usually characterize meltout tills (Boulton 1971; Lawson 1979b, 1989; Dreimanis 1989; Ronnert 1992). The phenoclast fabric of each diamicton slab could differ from that of the ice that reincorporated the diamicton slab. If the diamicton masses were not transported far, reorientation of the phenoclasts within the diamicton slabs may have been limited. Englacial transport would also have limited the intensity of phenoclast reorientation. The unit therefore would be composed of an unknown number of diamicton slabs, each with its own clast fabric, that were brought together during deposition.

A second mechanism to account for the low S1 eigenvalues and lack of consistent orientation is that the diamicton had undergone internal remobilization, as described in Rusczyñska-Szenajach (1983). Thin-sections from Lyons Creek and Battle Creek sections containing contacts between diamicton and bedrock blocks often show intrusion of diamicton along the joint faces of the blocks, especially along the margins (Chapter 5, this volume). These intrusions along the margins of the shale blocks indicate that in some areas the diamicton was highly water saturated during meltout. Sufficiently high pore pressures were generated to inject diamicton stringers along joints into the shale blocks. This over-saturated state must have been localized, since pervasive remobilization would have destroyed the thin stringers, bedrock slabs, and sand lenses but internal remobilization coupled with incorporation of debris slabs would explain the low S_1 eigenvalues and multimodal nature of the fabrics.

Incorporation of older slabs of diamicton may also explain the presence of numerous oblong and flat-iron shaped striated shale phenoclasts. Flat-iron shaped phenoclasts are most often associated with lodgement (Boulton 1978), but can be found in meltout tills (Dreimanis 1989). If these phenoclasts were derived from previously deposited diamicton masses, their shapes would reflect their initial mechanism of formation or deposition.

The shapes of the shale phenoclasts may reflect break up of shale blocks during transport. A continuum is observed in the sections from intact shale blocks, to blocks with thin diamicton injections along joint surfaces, to those in which the diamicton injections have begun to disaggregate the block. The final stage would be complete disaggregation into shale pebbles encased in a diamicton matrix. The shale pebbles may have been derived from the rounded and oval sub-units observed in thin-sections from large intact shale blocks.

The quartzite phenoclasts encased within apparently intact shale blocks are problematic and cannot be explained solely through the mechanisms of passive meltout. The position of these phenoclasts must also reflect the transport history of the block. Goldthwait (1968) has shown that a lateral down-ice dispersion over a 20^0 arc occurs during ice transport of material eroded from small distinct outcrops. If, during transport, a large shale block underwent only the initial stages of dispersion, it could be partially disaggregated. This partial disaggregation would be accompanied by the influx into the shale blocks of thin ice bands, some containing quartzite phenoclasts. If the ice mass stagnated at this time, and the intruding ice bands had subsequently melted, the phenoclasts would become encased within the shale block (Fig 2.9). The reconsolidation of the blocks would be aided by the soft plastic nature of the shale and overburden pressure. The irregular iron-stained joint network along the margins of the large shale block in BC1 may be the planes along which the block disaggregated and subsequently recombined. The diamicton stringers along the basal and lateral margins of the block would be formed by intrusion of water-saturated diamicton along cracks during reformation of the block. This style of breakup and recombination of the bedrock blocks reinforces the interpretation that the blocks were not emplaced by lodgement or other disruptive process but must be released by a passive process such as meltout. It also indicates that the blocks did not undergo long-distance basal transport.

Unit 2. Discontinuous stratified sediments interlayered with diamicton.

Unit 2 is less complex than Unit 1. It consists of well-sorted sand and silt beds (usually more than 50% of the volume of the unit) intercalated with diamicton layers of irregular thickness (Fig.2.3). The base of the first thick continuous sand bed was established as the lower limit of Unit 2. This base is very irregular, rising or falling up to 2 m over short lateral distances (2 to 3 m). The diamicton layers are commonly 10 to 100 cm thick, have sharp but irregular upper and lower contacts, and are indistinguishable in color, texture, and pebble lithology from the diamicton of Unit 1 (Fig. 2.13). The bulk texture is 25% sand, 36% silt, and 39% clay (see Appendix 3). Diamicton layers are generally discontinuous but some can be traced several tens of meters through Unit 2. The diamicton layers can have sub-horizontal orientations but more commonly show some degree of folding and may be very contorted and disrupted. Some appear to have settled into underlying sorted sediment layers while some of the thinner diamicton interbeds are boudinaged. Small shale slabs, up to 130 cm in length and 15 cm thick, within deformed diamicton beds are similarly folded and contorted. Diamicton diapirs with no detectable preferred orientation extend from Unit 1 into Unit 2 cross cutting sand and silt beds. Phenoclast fabrics measured, in thicker diamicton layers in BC1, LC1 and LC2, were multimodal, had S_1 eigenvalues of 0.46 to 0.63, and did not show a consistent orientation (Table 2.1).

The sorted sediment beds, ranging from clay and silt beds to gravel lenses, that are interlayered with the diamicton layers, form the bulk of Unit 2. Planar crossbeds predominate but ripples and rare trough cross beds are present. In LC2 and BC1, channel-shaped lenses of fine and medium sand are surrounded by clay-rich diamicton. No diamicton layers are contained within these lenses. Tabular lenses, 0.7 m thick and 2 m long, composed of rounded to sub-rounded shale pebbles deposited in planar cross-beds were also observed. Shale pebble beds often occur as isolated units in larger sand bodies. Small-scale normal faults (5 to 30 cm offsets) that crosscut contorted sand beds, diamicton layers, and diapirs occur in most of the sand lenses. In some places the faults can be traced into the underlying Unit 1 (Fig 2.14).

In BC1, the sand and diamicton layers are complexly folded, contorted, and overturned (Fig. 2.2, 2.14, 2.15). The contortions are up to 30 m in length and 1 to 5 m in thickness. Some layers within the unit are complexly folded and overturned while others at the same level along the section are undeformed. The deformations commonly destroy any primary bedding structures within the sand and silt. Where deformation is not as pervasive, sedimentary structures such as fine laminations and ripples, planar cross beds, and tabular beds of unbedded fine sand and silt are detectable.

Genesis of Unit 2

Unit 2 is interpreted as a subaerial glacigenic debris-flow complex in which debris-flows (diamicton layers) are intercalated with stratified sediment (deposited by subaerial glacial streams). The generally tabular geometry of the diamicton beds in the unit and their irregular upper surfaces match descriptions of subaerial debris-flows

(Hubert and Filipov 1986; Ghibaudo 1992) and supraglacial debris flows (Hartshorn 1958; Lawson 1979a; Dreimanis 1989). The rapid lateral and vertical changes from diamicton to sorted stratified sediments in Unit 2 are also characteristic of subaerial glacially-derived debris-flow complexes (Boulton 1968; Boulton and Eyles 1979; Eyles 1979, 1983; Eyles et al. 1982). The shale pebbles are soft and fractured. This would ensure their rapid break-up. The shale pebble beds therefore were derived from extensive reworking of entrained local shale bedrock blocks and phenoclasts observed throughout Unit 1. The streams therefore did not transport the shale pebbles far. The associated sand, silt, and gravel beds were also locally derived. Weak, multimodal phenoclast fabrics, lacking consistent orientations, are commonly measured in debris-flow deposits (Lawson 1979b; 1981; Rappol 1985; Dowdeswell and Sharp 1986). The clayey diamicton layers are readily interpreted as glacigenic debris flows.

The prominent deformations observed at BC1 are large-scale soft sediment deformation features similar to structures described by Stone (1976), and Visser et al. (1984). These formed when saturated debris accumulations were suddenly loaded by thick debris-flow or sorted sediment layers. This caused the pore pressure within the loaded layers to increase and the effective pressure within them was reduced. The layers became unstable and were prone to movement and failure. Mobilization of the sediment caused failure, disruption, and foundering of the overlying sediment. This deformed, and in places, destroyed the primary sedimentary structures within the layers. The diamicton layers within the unit may have acted as impermeable barriers facilitating the increase in pore pressure until remobilization and diapir injection occurred. Similar features have been reported by Shaw (1983, 1987) and Broster and Clague (1987). The lack of a common orientation to the deformations and the presence of undeformed stratified beds in close association with deformed layers (less than 10 m separation) indicates that the over-pressured areas were highly localized.

The name of the unit reflects the complexity of its origin. Subaerial is used rather than supraglacial as there is no evidence that the sediment comprising the unit was supraglacial or was from lower in the ice and subsequently released subaerially. The debris-flows could be called flow tills but the degree of remobilization and movement shown by the diamicton layers in the unit has removed any feature inherited from the ice. Since the unit is an assemblage of sorted sand and silt intermixed with diamicton layers it is polygenetic and the term complex incorporates this. The resulting name therefore supplies a complete picture of the origin of the unit.

The relationship between Units 1 and 2

Diamicton diapirs, extending from Unit 1 into Unit 2 and crosscutting the stratification in the sand lenses of Unit 2, probably followed planes of weakness in the overlying unit. Their injection from Unit 1 into Unit 2 demonstrates that Unit 1 was still saturated and unconsolidated when Unit 2 was emplaced over it. The normal faults that extend from Unit 1 into Unit 2 and cross-cut sorted sediment and diamicton diapirs are similar to those described in McDonald and Shilts (1975) and indicate that underlying buried ice blocks were present until after deposition and partial consolidation of Unit 2. Both the faults and diapirs show that the two units can be considered essentially contemporaneous and that meltout of buried ice continued after deposition of Unit 2. Persistence of buried ice beneath a thick debris blanket is well documented (Sharp 1949; Clayton 1964; MacKenzie 1969; Driscoll 1980). Its presence beneath Unit 2 supports the subglacial meltout interpretation of Unit 1.

The irregular boundary between Units 1 and 2 may reflect the lowering of Unit 2 onto Unit 1 following the meltout of buried ice masses. Some of the dragfolding and other disruptions in the strata of Unit 2 likely reflect this lowering.

Depositional summary

Units 1 and 2 are members of a continuum varying from meltout till to glaciofluvially reworked till. The distribution of each unit is dependent upon the relative dominance of each depositional process. Unit 1 extends to the surface with its meltout characteristics intact where it was deposited in a stable location, dewatered sufficiently to prevent remobilization and not reworked by glacial streams. Random variations in dewatering, stability of accumulated debris, and loci of sorted sediment deposition ensures that the Unit 2 is not uniform in appearance or distribution.

The difference in character between Unit 1 and 2 not only reflects the greater reworking of sediment near the surface of the ice mass but reflects the marked difference in debris volume between the basal debris-choked ice and ice above this where the debris was more disseminated. Stacking of debris-laden ice, as indicated by the bedrock slabs, may have increased the overall thickness of the debris-charged basal zone. The sediment of Unit 2 should contain a large component of basal or near basal debris. Substantial thicknesses of overlying relatively debris-free ice probably melted before exposure of ice containing sufficient debris to form a recognizable deposit.

Above the debris-rich basal zone, the debris volume and its distribution are expected to have been more irregular, and the melt rate to have been more rapid due to increased heat influx, creating large volumes of meltwater. This meltwater would have penetrated deeper into the ice along cracks, crevasses, and moulins. The increased meltwater flow in the upper part of the ice would have facilitated sediment reworking, generating large volumes of sorted stratified sediment. Large debris-flows could divert streams, changing the locations of reworking and deposition. Following such random shifts in stream path, sorted sediments were deposited next to or on top of unstable water-saturated diamicton layers. The diamicton accumulations were then remobilized and in some areas subject to diapirism. Thus glacially-derived debris-flows were intermixed with glaciofluvial sediment. The result was a complex mix of intercalated sand lenses and debris-flows in which individual layers are discontinuous and rapidly pinch-out. Rapid sediment deposition from supraglacial streams and debris-flow deposition, locally over pressured saturated diamicton and sorted sediment layers initiating diapirism and soft-sediment deformation.

Unit 2 also acted as an insulating debris blanket for the underlying debris-laden ice. This decreased the melt rate of the buried ice, helping to ensure that the meltwater generated was removed and that pore-water pressure remained low. The underlying subglacial meltout till was therefore not affected by widespread remobilization and its englacial features were well preserved. The weight of the overlying ice and debris, consolidated the sediment as it was released forming a dense compact deposit.

Discussion

A. Limitations on deposition of a thick (>10m) areally extensive meltout till

The two units described are distributed throughout the study area (Fig. 2.1) with only the proportions of each unit differing from section to section. Lawson (1979a) observed that 85% of the glacially-derived diamicton deposited at the terminus of the Matanuska Glacier in Alaska, was resedimented or substantially reworked by meltwater (therefore there was little preservation of till 'sensu stricto'). Paul and Eyles (1990) stated that a subglacial meltout till cannot be preserved unless a set of restrictive preconditions is met. Deposition of an areally extensive thick meltout till therefore requires explanation.

The first aspect of the subglacial meltout till to be examined is its thickness. The till ranges from 5 to 35 m in thickness. This would have required a dense concentration of debris near the base of the ice. The shale of the area is soft and easily eroded. In many places in the till, shale pellets make up as much as 35% of the diamicton's total volume, and shale slabs and blocks are common throughout Unit 1. The slight color difference between the shale blocks and the encasing diamicton matrix indicates that the matrix is probably largely composed of comminuted shale. The local shale therefore contributed greatly to the diamicton.

Simple entrainment of easily eroded bedrock does not explain the thickness of the diamicton in the sections. Stacking of debris-charged ice layers derived from near the base of the ice and containing diamicton blocks and bedrock slabs, substantially increased the volume of debris in the ice. Compressive flow conditions, generated as the ice ascended the west slope of the Old Man On His Back Plateau, or freeze-on processes (Clayton and Moran 1974; Bluemle and Clayton 1984) would facilitate this stacking. The multiple bedrock slabs in BC1 demonstrate that stacking within the ice did occur. A highly charged basal debris layer is therefore probable.

The second requirement is that the debris-charged basal ice be melted at a low enough rate that the meltwater produced is able to drain from the clay-rich sediment. The deposit therefore consolidates as it is deposited preventing it from becoming over-saturated and prone to remobilization (Rusczyñska-Szenajach 1983). Lawson (1979a, 1989) describes formation of such dense, consolidated subglacial meltout till. Increasing the debris concentration in the ice reduces the water volume produced as the dirty ice melts, helping prevent oversaturation and remobilization. The major controls on the rate of melting of stagnant ice are geothermal heat inflow, the infiltration of surface heat, and influx of warmer surface meltwater.

Geothermal heat is the smallest component of the total heat inflow. Paul and Eyles (1990) calculated that the geothermal heat influx in the Prairies is incapable of melting sufficient basal ice to create unstable saturated conditions in the sediment. Their modeling indicates that even with additional heat from meltwater flowing in englacial tunnels, the rate of melting was incapable of producing unstable saturated basal conditions. Only in areas with direct atmospheric contact, was the melting sufficiently rapid to create a saturated subglacial deposit. Surface meltwater input and direct atmospheric contact are therefore needed to generate a rate of ablation high enough to cause saturation and promote instability.

Infiltration of warm surface meltwater along temporary meltwater conduits and surface heat inflow from direct atmospheric contact in crevasses, moulins, and open basal caverns, supply the bulk of the heat used to melt a stagnant ice block (Paul and Eyles 1990). During the initial stages of melting, the ablation rate is high and the ice surface is irregular. As debris accumulates on top of the ice the rate of ablation slows. Even a thin sediment cover dramatically reduces surface heat infiltration (Sharp 1949; Ostrem 1959; MacKenzie 1969; Mickelson 1973; Driscoll 1980; Watson 1980). McKenzie estimated that ablation beneath a thick debris cover would be about 0.6 mm per day. Preservation of ice beneath a thick sediment cover is demonstrated by the drunken forests on the terminal parts of the Malaspina and Klutan Glaciers (Driscoll 1980; Watson 1980). Clayton (1967) estimated that ice buried beneath debris on the Missouri Coteau took about 3,000 years to melt completely.

The formation of an insulating debris-blanket not only reduces heat penetration but also acts as a barrier to erosion and reworking by surface meltwater streams. Winnowing and reworking is therefore restricted to the upper part of the debris layer. Unit 2's major role therefore, was its ability to insulate the debris-laden ice below from both heat infiltration and glaciofluvial reworking. Faults and dragfolds in the sorted stratified sediment of Unit 2 indicate that basal (and perhaps englacial) ice was preserved for a time sufficient for Unit 2 to become partially consolidated.

The maintenance of the debris mantle is critical to the preservation of an intact subglacial meltout deposit (Paul and Eyles 1990). Sediment accumulating on top of stagnant ice masses is usually assumed to undergo several stages of relief inversion because of the irregular surface topography and the non-uniform debris distribution (Gravenor and Kupsch 1959; Paul and Eyles 1990). No evidence for multiple relief inversions was detected. It may be that relief inversion was confined to Unit 2 and did not affect the ice mass underneath. The normal faults in Unit's 1 and 2 are evidence of consolidation and brittle fracture. As brittle materials do not flow, any relief inversion that occurred must have taken place early in Unit 2's genesis. The author has observed the same facies sequence in adjoining hummock and hollow pairs in hummocky terrain in the Wetaskiwin area of Alberta. The facies sequence in the hollow was displaced downwards 5 to 8 metres. This suggests that the sedimentary units in the hollow had been passively lowered into place and had not undergone inversion (Kulig 1985). The same inversion-free lowering may have occurred in the study area.

Paul and Eyles (1990) believed that subglacial meltout till is preferentially preserved on the proximal side of ice-cored ridges formed by thrusting at a margin undergoing incremental stagnation. Debris is therefore distributed in highly irregular zones. Topographic differences between thrusted ridges also create major irregularities in the distribution of debris-laden ice. As the debris-laden ice melts, remobilization of debris occurs, reducing these topographic and distributional irregularities. If the saturated debris was more evenly distributed and it was deposited in a stable setting there would be less reworking and better preservation of the subglacial meltout till. Ice blocks or stable sediment masses around the unstable saturated debris could hold the unstable debris in place. It would be incapable of movement and could consolidate slowly without failure.

Thrusting and incremental stagnation create localized discontinuities in sediment distribution and topography that result in remobilization of debris. Arcuate ridges and terminal, or recessional moraines, commonly associated with this type of ice margin retreat, are not present in the area. The study area appears to have been affected by broad regional, not incremental, stagnation. The ice passed over the Comrey preglacial bedrock high (Fig. 1) and ascended onto the Old Man On His Back Plateau to the east when it entered the area. The compressive flow conditions caused by the ascent would have facilitated stacking of bedrock slabs (sandstone and shale) in the ice. During wastage, the debris-charged ice between these bedrock highs would have been the first to become isolated and inactive. Regional stagnation of ice, with a concentrated debris-zone at or near its base therefore contributed to formation of an areally extensive subglacial meltout till.

The thick, extensive subglacial meltout till of the study area required a debris-charged basal debris layer, a low rate of melting that reflects a low input of geothermal heat, and the presence of an insulating debris mantle. The regional stagnation of the debris-laden ice creating a stable setting for meltout also aided preservation of the till. An additional factor, the direction of proglacial drainage, was also important.

B. Influence of proglacial drainage on till deposition

Unusually thick (greater than 10 m) and extensive subglacial meltout till has not been recognized in the extensive river sections exposed along the Old Man or South Saskatchewan rivers (Proudfoot 1985). Nor have they been reported from other parts of the Prairies. The low probability of obtaining the necessary combination of features (the dense concentration of debris, formation of a continuous and persistent debris mantle, low melt rate, and stable deposition setting) may account for this. The preservation of thick meltout till may also reflect the ability of the ice margin to drain freely.

North of the Cypress Hills and over most of the Prairies, downslope ice retreat blocked meltwater drainage. Numerous large ice-marginal lakes were therefore impounded (Stalker 1960, 1973; Christiansen 1968, 1979, 1992; St. Onge 1972; Shetsen 1984, Fig. 19; Kulig 1985; Proudfoot 1985; Dyke and Prest 1987). Unstable debris aprons were deposited in these lakes near the ice-water contact from rainout (Barnett and Holdsworth 1974; Gibbard 1980) and subaerial and subaquatic glacially-derived, debris-flows (Evenson et al. 1977; Kulig 1985; Proudfoot 1985). Remobilization of these debris aprons emplaced debris farther into the lake. Intercalated with the debris-flow and rain-out sediment were silt and clay couplets, sand and silt laminae from underflows, and sand lags from current winnowing of the diamicton surfaces. Delta deposits at the entry points of streams (Gibbard 1980; Evenson et al. 1977; Kulig 1985) are also common. Dropstones released from icebergs are also a ubiquitous feature (Ovenshine, 1970). Such glaciolacustrine deposits have been recognized throughout Alberta and make up a significant proportion of the glacigenic sediment. These deposits commonly have gradational contacts with underlying glacigenic units and are often called stratified diamictons (Proudfoot 1985; Odynsky et al. 1950, 1952; May 1977; Shaw 1987; Kulig 1985).

On the south slopes of the Cypress Hills and Milk River Ridge, drainage is to the south and southeast. When the Late Wisconsinan ice entered the area from the northwest, it did not disrupt the regional drainage pattern and unrestricted drainage to the southeast along the ice margin was maintained. The maintenance of free drainage prevented the formation of large ice-marginal lakes and deposition of the sediment assemblages associated with them. Instead a thick, extensive sequence of meltout till overlain by a subaerial glacigenic debris-flow complex, as described, was deposited.

The two settings are fundamentally different in their depositional processes. Two distinct glacigenic depositional models are therefore required for the Prairies. Where free drainage is maintained, deposition is dominated by meltout and subaerial glacially-derived debris-flow complexes, with localized contributions from lodgement and glaciotectonism. The proportions of each till type reflect local conditions within the ice mass. Where drainage is ponded and ice marginal lakes form, glaciolacustrine depositional processes predominate with minor contributions from other glacial processes such as meltout and lodgement (Shaw 1982,1987; Kulig 1985).

C. Regional stratigraphy

Evidence, such as buried channels or multiple till sections with erosional or glaciofluvial contacts indicating that more than one glaciation affected the study area, was not observed in any of the sections. Units 1 and 2 were deposited during the same glacial event as shown by the diapirs, faults and irregular indistinct contact between the two units. The lithostratigraphic equivalents of these units outside the study area have not yet been established. The units are temporally equivalent to the Battleford Formation (Christiansen 1968, 1992) but they are thicker and differ substantially from the characteristics given for the Battleford Till (Christiansen, 1968, 1992). They were deposited during the Late Wisconsinan Lost Wood Glaciation (Fenton 1984). It is uncertain whether any of the unnamed multiple tills deposited at Medicine Hat during the last part of the Wisconsin Glaciation (Stalker 1977; Proudfoot 1985) can be correlated to the units investigated. Since Units 1 and 2 extend into northern Montana they should also correlate to the Fort Assiniboine Till (Fullerton and Colton 1986), the Late Wisconsinan surface till that covers northwest Montana. The Middle Creek Advance, the second most extensive of the advances described for the West Lobe (Chapter 6, this volume), did not reach the study area. The sediment therefore was deposited during the Underdahl Advance, the most extensive of the Late Wisconsin glacial events in the region.

Conclusions

The two glacial sedimentary units observed in the study area are composed of a subglacial meltout till and an overlying subaerial glacigenic debris-flow complex. The widespread distribution of the thick subglacial meltout till is unusual, but the preconditions necessary for its deposition and preservation (a very large basal and englacial debris concentration, slow meltout beneath a protective debris mantle, regional stagnation) were present. The maintenance of drainage away from the ice margin, which precluded the impounding of large ice contact lakes and deposition of subaqueous glacigenic sediment assemblages (Evenson et al. 1976; Gibbard 1980; Kulig 1985; Proudfoot 1985), also contributed greatly to the deposition of the thick, areally-extensive meltout till.

Two broad depositional environments may be present on the Prairies: one in which the ice margin is freely drained and deposition consists of an assortment of subglacial and supraglacial till varieties; the other where impounded regional drainage generates large ice-contact lakes in which subaquatic depositional processes dominate. Reducing glacial sedimentation to two broad depositional environments is simplistic but provides a framework for further investigations.

Table 2.1. Statistics for phenoclast fabrics measured in the units in the Battle and Lyons Creek sections.

FABRIC NO.	UNIT	TREND	PLUNGE	S_1	S_2	MULTI-MODAL	CLAST NO.
BC1-1	1	48.4	13.7	0.62	0.29	Y	30
BC1-2	1	282.2	3.5	0.55	0.28	Y	25
BC1-3	1	48.2	6.8	0.65	0.24	Y	30
BC1-4	1	277.6	30.6	0.61	0.30	Y	30
BC1-5	1	5.4	12.8	0.54	0.40	Y	30
BC1-6	1	305.9	22.3	0.77	0.15	N	30
BC1-7	1	288.1	30.0	0.58	0.34	Y	30
BC1-8	2	320.7	1.8	0.58	0.31	Y	25
BC1-9	2	123.1	2.2	0.58	0.32	Y	30
LC1-1	1	90.6	10.0	0.63	0.24	Y	30
LC1-2	1	77.7	10.7	0.59	0.28	Y	26
LC1-3	1	267.3	36.6	0.64	0.29	Y	25
LC1-4	1	34.5	2.6	0.73	0.17	Y	25
LC1-5	1	264.4	22.8	0.70	0.14	Y	25
LC1-6	1	123.6	8.7	0.71	0.24	Y	25
LC1-7	1	268.9	35.8	0.53	0.33	Y	25
LC1-8	2	242.8	0.4	0.53	0.38	Y	25
LC1-9	2	248.9	60.4	0.45	0.32	Y	25
LC2-1	1	103.3	7.7	0.48	0.32	Y	30
LC2-2	1	284.5	22.0	0.63	0.23	Y	31
LC2-3	1	226.1	7.9	0.59	0.33	Y	30
LC2-4	1	334.0	13.1	0.66	0.21	Y	30
LC2-5	1	82.5	3.2	0.59	0.28	Y	30
LC2-6	1	288.3	9.4	0.58	0.29	Y	25
LC2-7	2	273.8	5.6	0.42	0.39	Y	27

Figure 2.1. Location of the study area (box) and its relationship to the major features of the area. The Cypress Hills have been divided into three blocks, the West Block (WB) straddling the Alberta/Saskatchewan Border, the Center Block (CB), the smallest of the blocks, and the East Block (EB). OM Old Man On His Back Plateau, WMU Wood Mountain Upland, BAC - Battle Creek, BC1,2 - Battle Creek sections, FC Frenchman Channel, LC1,2,3 - Lyons Creek sections, CL Climax, EL Elkwater, EE Eastend, MC Maple Creek, RB Robsart, SH Shaunavon.

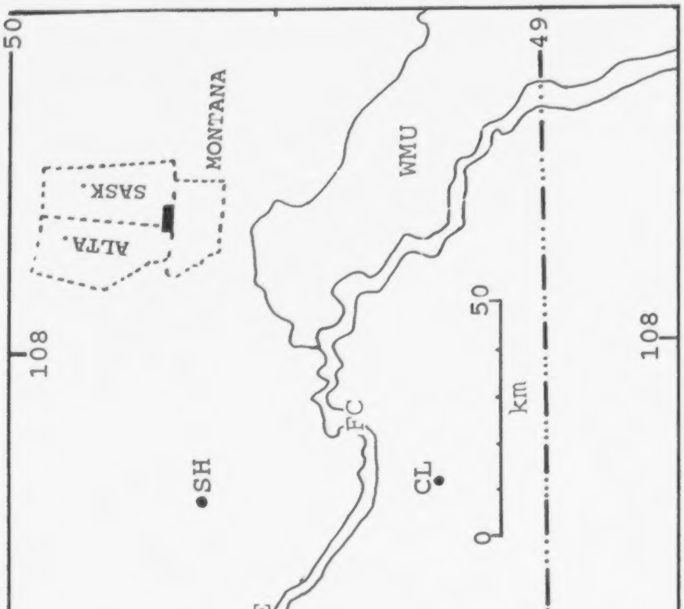

50
108
109
50
110
108
109
49
49
ALTA.
SASK.
MONTANA
WMU
FC
SH
CL
0
50
km
EE
EB
Hills
FC
MC
CB
RB
BAC
OM
LC 1 2 3
BC1
BC2
Cypress
WB
EL
ALBERTA
SASKATCHEWAN
MONTANA

Figure 2.2. Location of sections along Battle and Lyons Creeks.

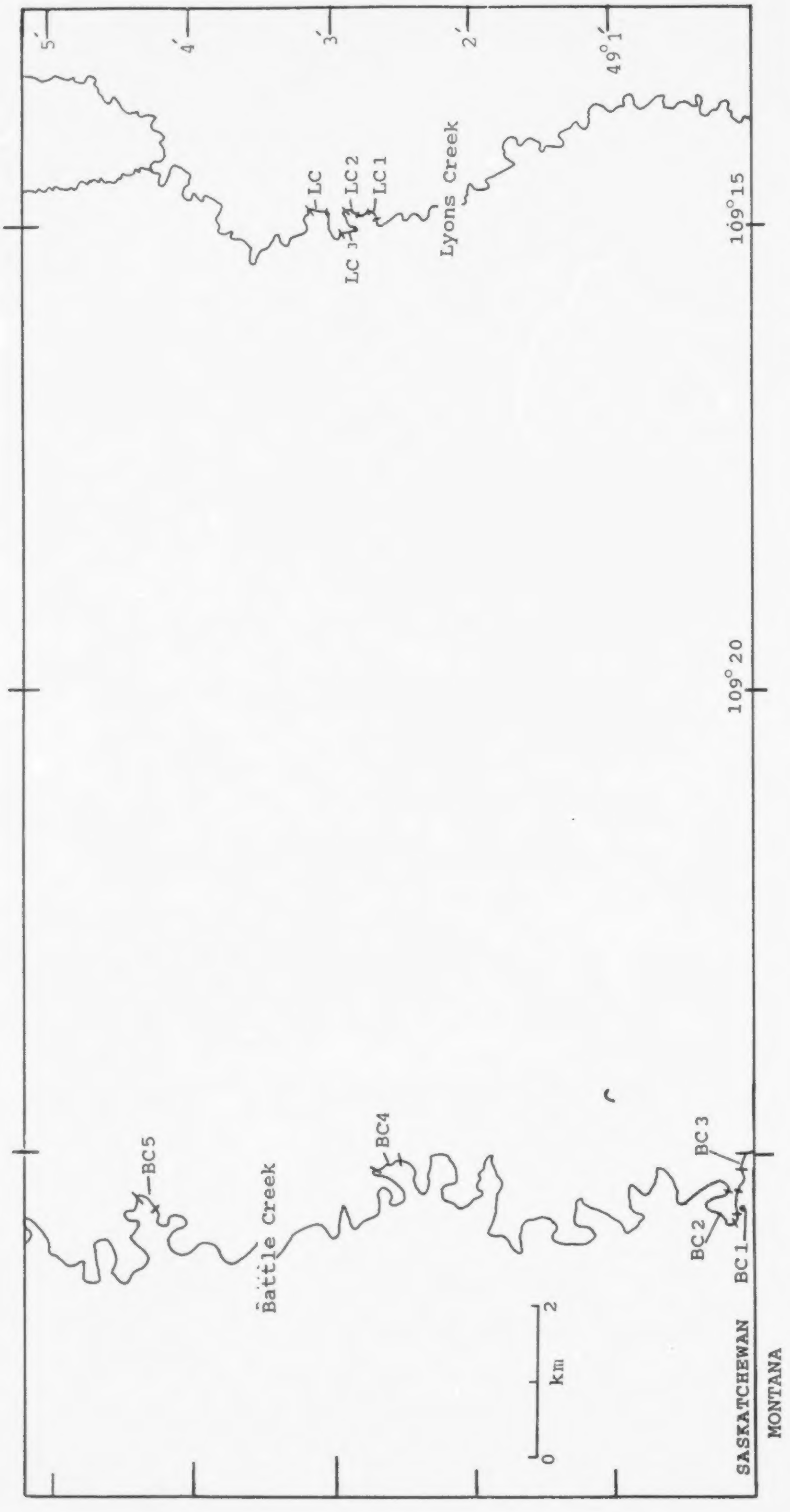
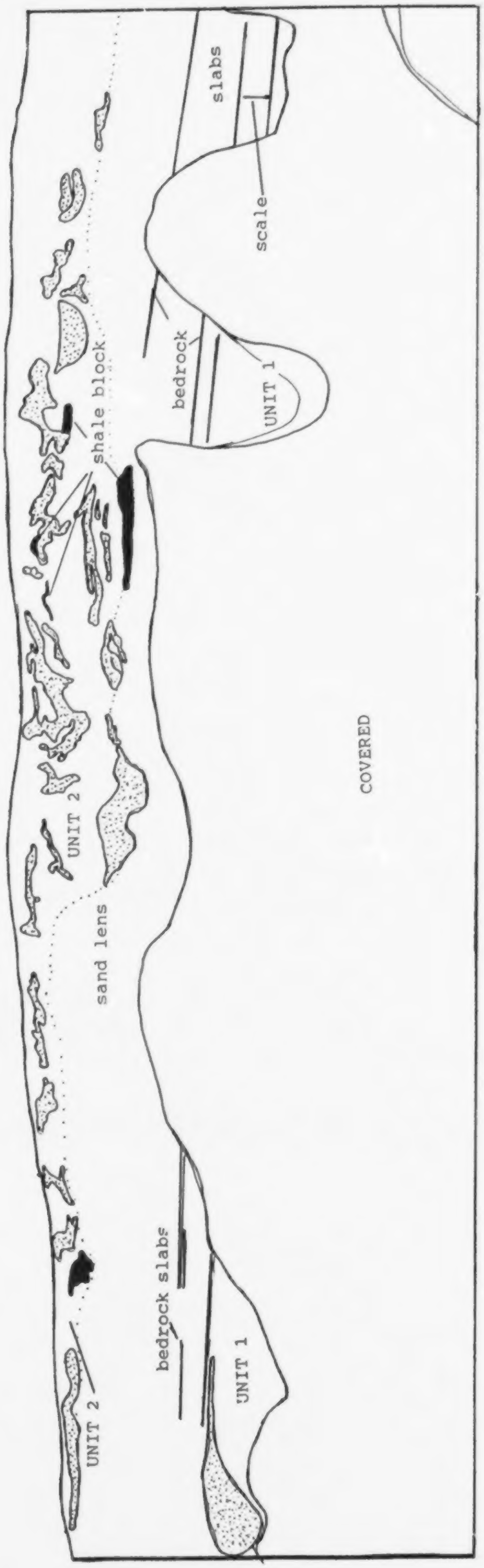

5'
4'
3'
2'
49°1'
Lyons Creek
LC
LC 2
LC 1
LC 3
109°15
109°20
Battle Creek
BC5
BC4
BC3
BC2
BC1
0
km
2
SASKATCHEWAN
MONTANA

Figure 2.3. Units 1 and 2 at section Battle Creek 1 (BC1) illustrating the relationships between the two units and the features they contain. The dotted line represents the approximate boundary between Units 1 and 2. The scale is a shovel 1.7 m high.


slabs
scale
bedrock
UNIT 1
shale block
COVERED
UNIT 2
sand lens
bedrock slabs
UNIT 1
UNIT 2

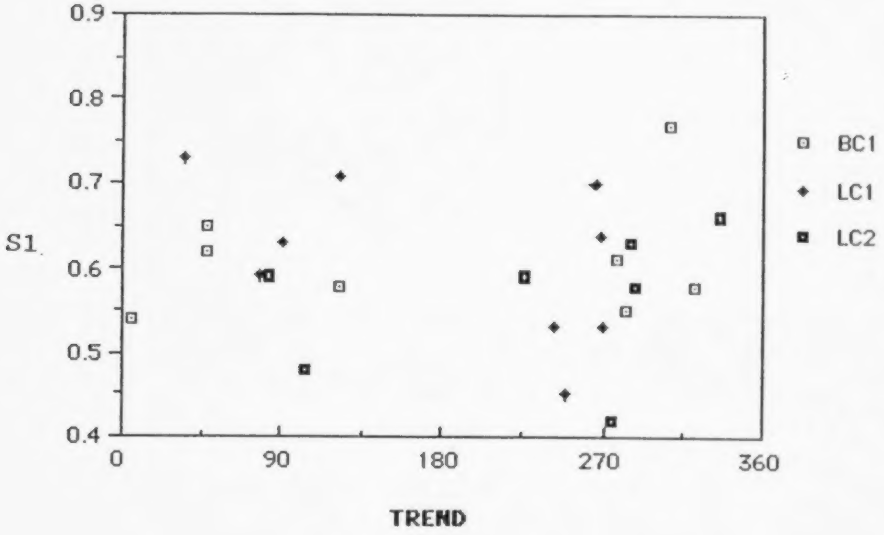


Figure 2.4. Plot of S_1 eigenvalue vs. Trend for phenoclast fabrics measured in Unit 1 of the sections showing clustering around 90^0 and 270^0.

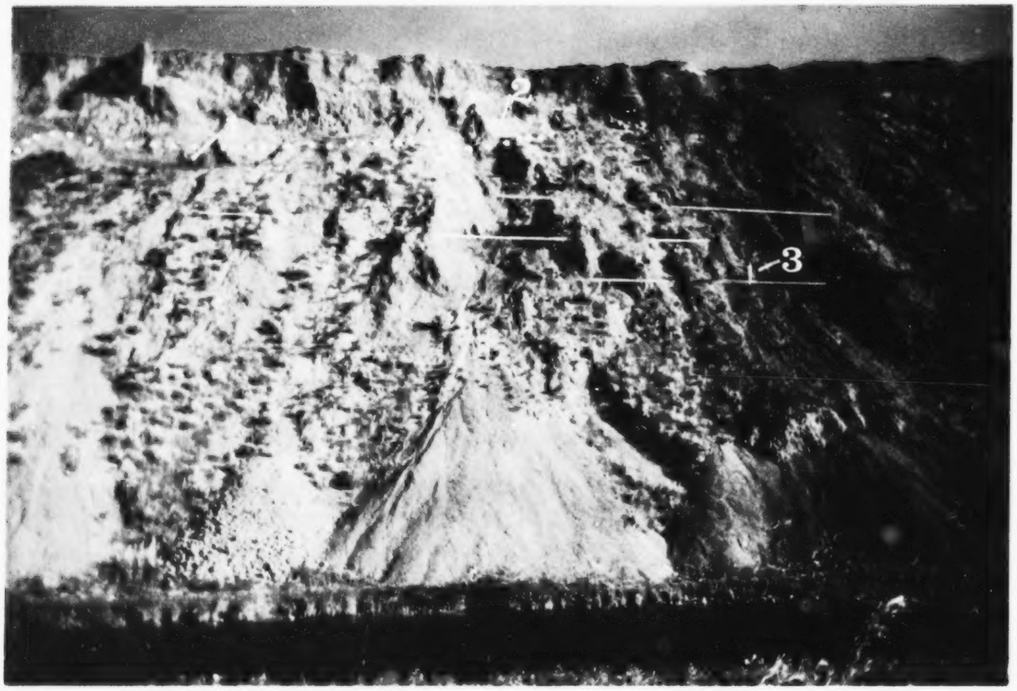


Figure 2.5. Units 1 and 2 at section BC1, a rivercut section along Battle Creek. The boundary between the units is highlighted by white dots. 1 is the large shale slab near the top of Unit 1. 2: is the channel-like lens in Unit 2. White lines show the location of the bedrock slabs. 3 is the 1.75 m shovel used for scale. The section is 32 m high.

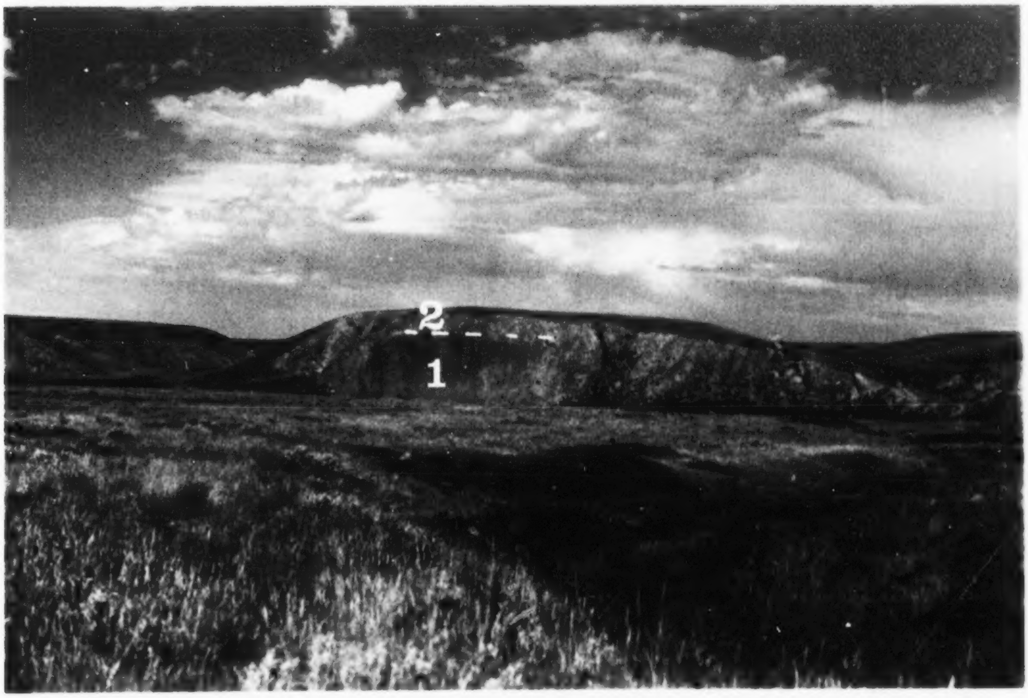


Figure 2.6. Lyons Creek Section 1 a rivercut section incised into the till plain. The unbedded meltout till (Unit 1) is overlain by interbedded sand and diamicton of Unit 2. Unit 2 has a recessive nature. The section is 15 m high.

Figure 2.7. Close-up of a green silty sandstone stringer in section BC1.

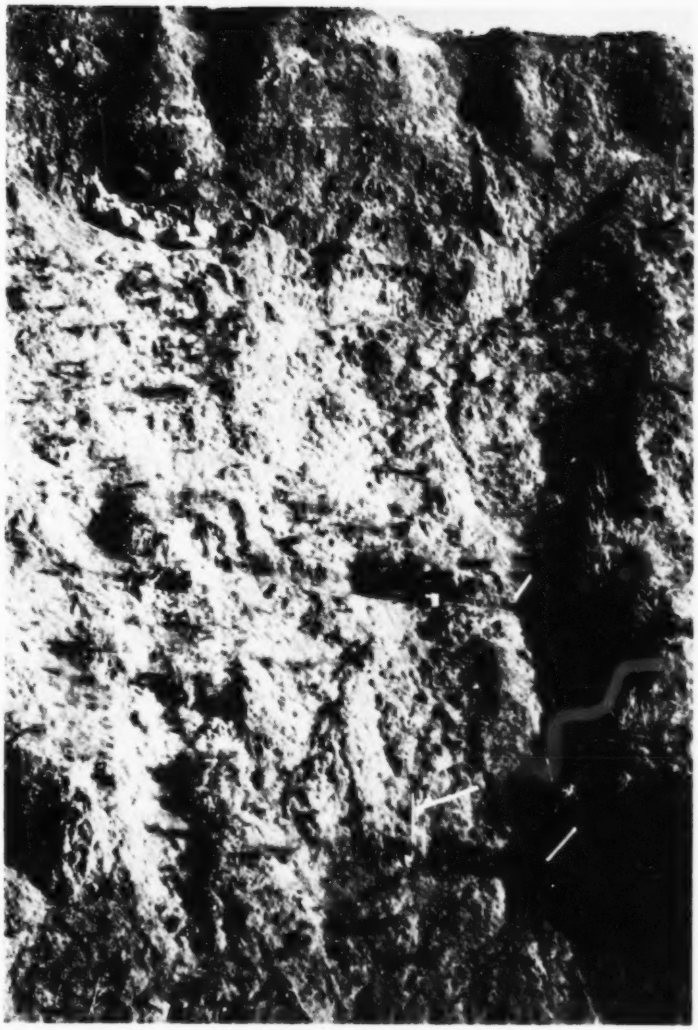


Figure 2.8. Close-up of two of the stacked sandstone slabs in section BC1. The shovel handle is 1.75 m..

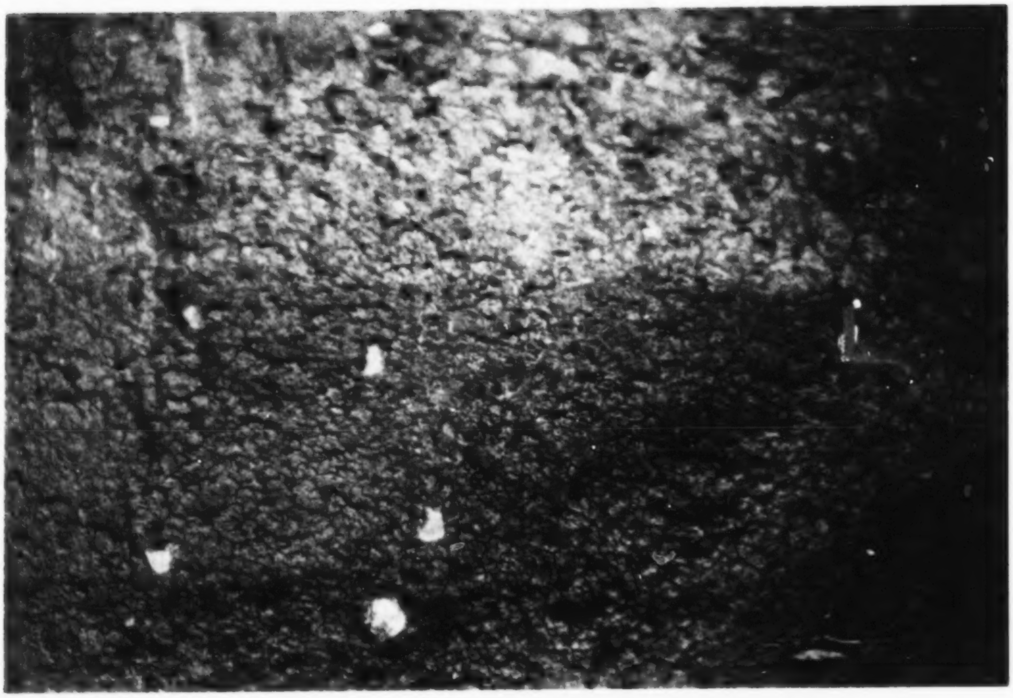


Figure 2.9. Quartzite phenoclasts enclosed within shale block near the top of Unit 1 in section BC1.

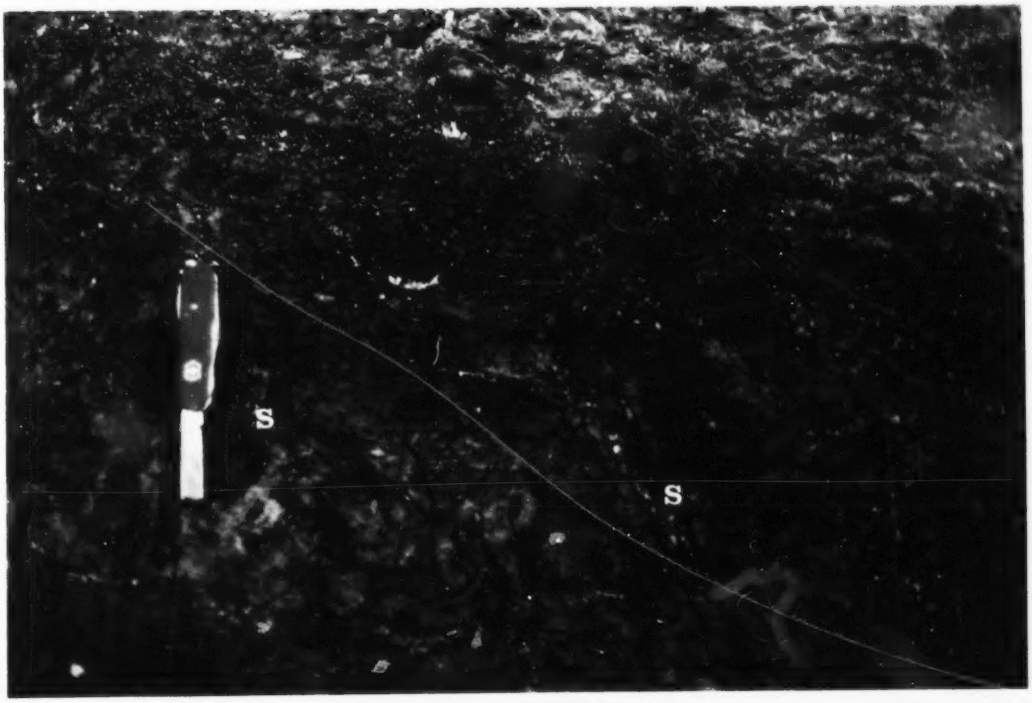


Figure 2.10. Diamicton stringers injected into the base of the large shale block shown in Figures 2.3, 2.5 and 2.9. The stringers are labeled S.

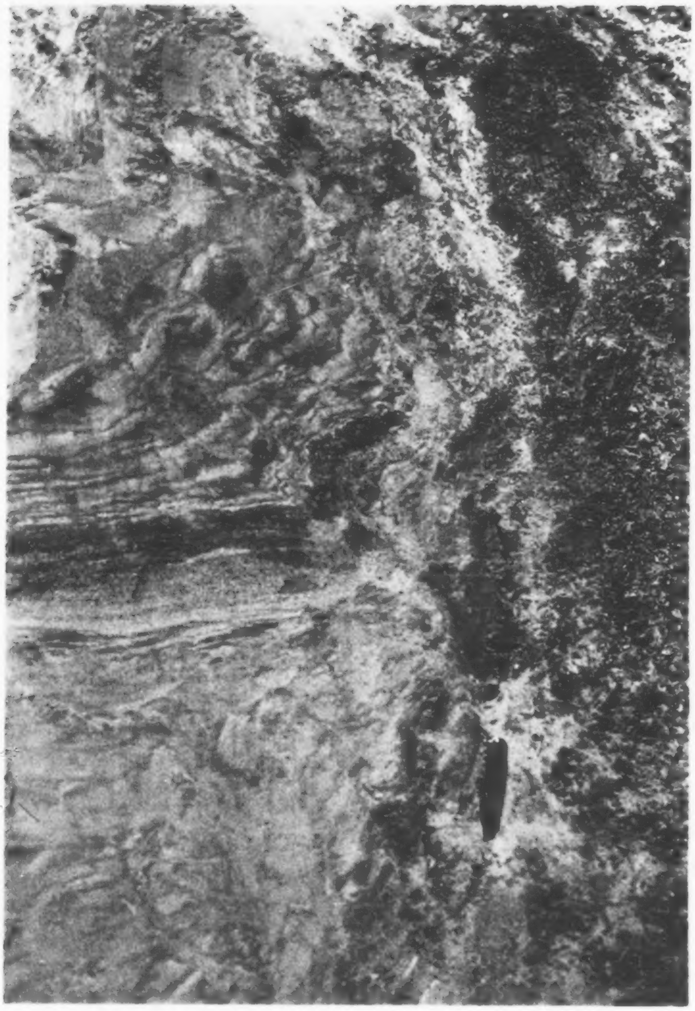


Figure 2.11. Deformation of sedimentary structures at the margin of a sand lens encased in diamicton in Unit 1 of section LC3.

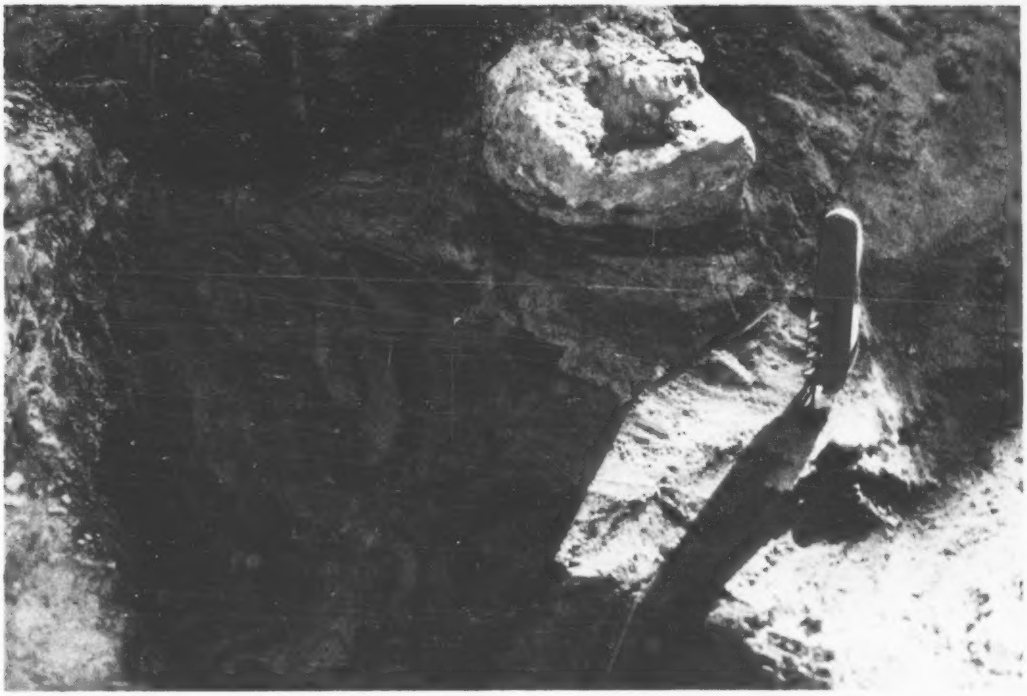


Figure 2.12. Stone held within diamicton and extending into the underlying sand lens. Photograph of Unit 1 of section LC2.

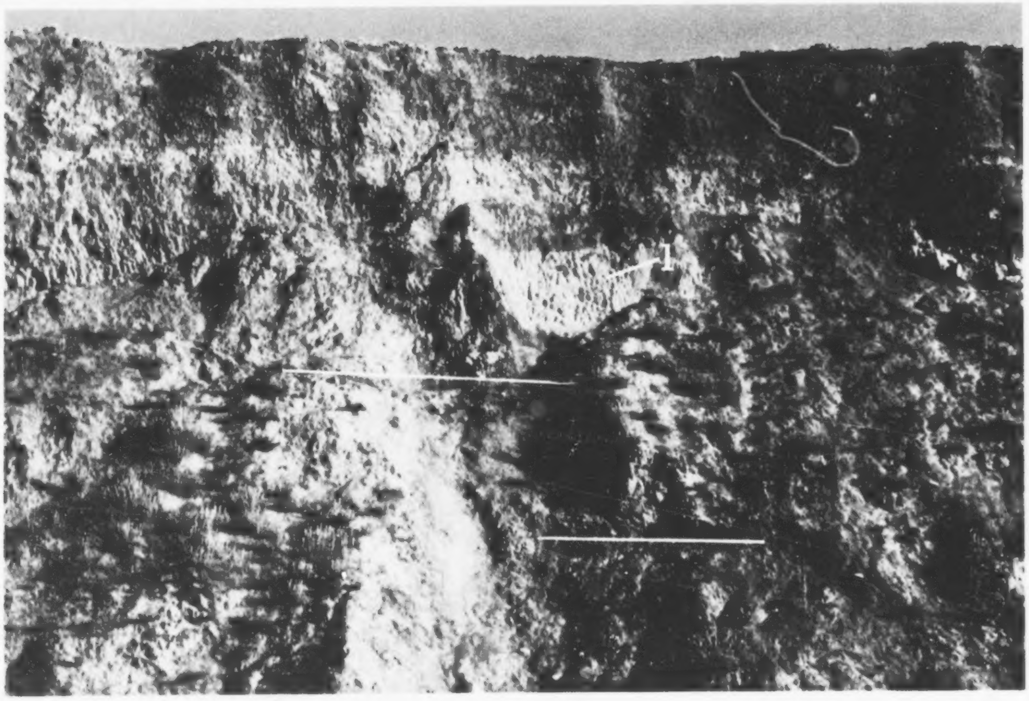


Figure 2.13. Undeformed sand lenses in section BC1. The 2 m high channel-shaped lens (1) is 20 m west of a contorted sand lens at the same level. The channel-shaped lens is also visible in Figures 2.3 and 2.5. Bedrock slabs are indicated by white lines.

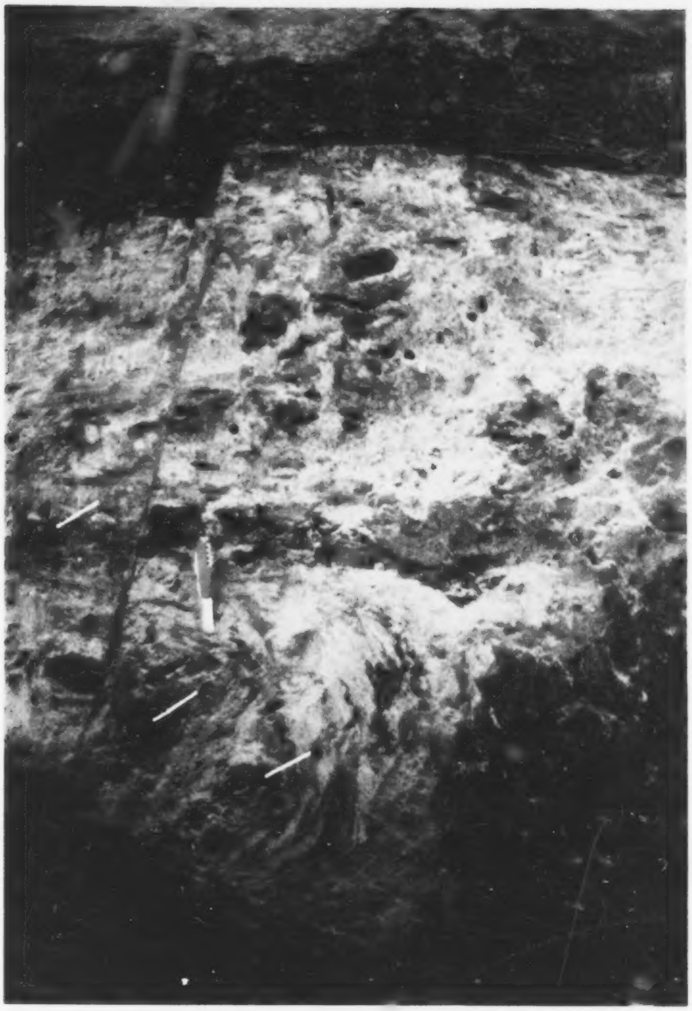


Figure 2.14. Normal faulting disrupting sands in a lens in Unit 2 of section BC1. Thin diamicton diapirs injected into the lenses (shown by white lines) are also offset. The fault extends from Unit 1 into Unit 2.

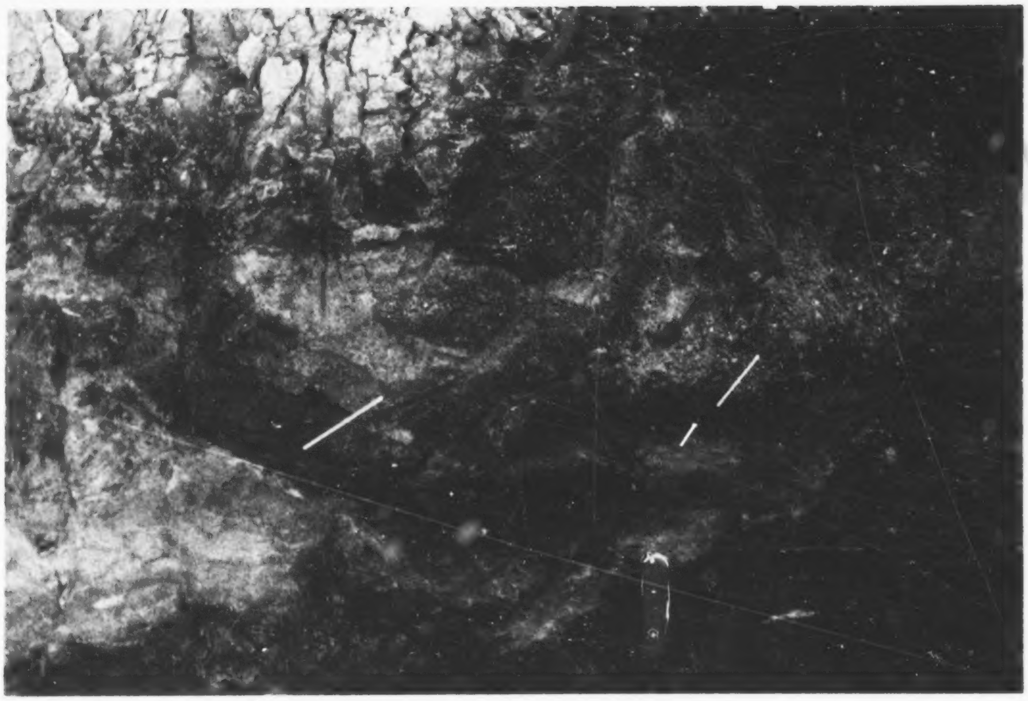


Figure 2.15. Diapirs (indicated by white lines) extending from Unit 1 into a sand lens in Unit 2 of section LC2.

References cited

Aber, J.S., 1985, The character of glaciotectonism: Geologie en Mijnbouw, v. 64, p. 389-395.

Banham, P. H., 1977, Glaciotectonites in till stratigraphy: Boreas, v. 6, p. 101-105.

Barnett, D.M., and Holdsworth, G., 1974, Origin, morphology and chronology of sublacustrine moraines, Generator Lake, Baffin Island, N.W.T.: Canada. Canadian Journal of Earth Sciences, v. 11, p. 380-408.

Bluemle, J.P. , and Clayton, Lee, 1984, Large-scale thrusting and related processes in North Dakota: Boreas, v. 13, p. 279-299.

Boulton, G.S., 1968, Flow tills and related deposits on some Vestspitsbergen glaciers: Journal of Glaciology, v. 7, p. 391-412.

Boulton, G.S., 1970, On the deposition of subglacial and meltout tills on the margin of certain Svalbard glaciers, Journal of Glaciology, v. 9, p. 231-245.

Boulton, G.S., 1971, Till genesis and fabric in Svalbard, Spitsbergen: *in* Goldthwait, R.P., ed., Till: a Symposium. p. 41-72,

Boulton, G.S. 1972, Modern Arctic glaciers as depositional models for former ice sheets: Quarterly Journal of the Geological Society of London, v. 128, p. 361-393,

Boulton, G.S., 1978, Boulder shapes and grain-size distribution of debris as indicators of transport paths through a glacier and till genesis: Sedimentology, v. 25, p. 773-779.

Boulton, G.S., and Eyles, N., 1979, Sedimentation by valley glaciers: a model and genetic classification: *in* Schultcher, J.M., ed., Moraines and Varves, Balkema Press, Rotterdam, p. 11-25.

Bretz, J.H. 1943. Keewatin end moraines in Alberta, Canada: Bulletin of the Geological Society of America, v. 54, p. 31-52.

Broster, B.E., and Clague, J.J., 1987, Advance and retreat glacigenic deformation at Williams Lake, British Columbia: Canadian Journal of Earth Sciences, v. 24, p. 1421-1430.

Broster, B.E., and Seaman, A.A. 1991, Glacigenic rafting of weathered granite: Charlie Lake, New Brunswick: Canadian Journal of Earth Sciences, v. 28, p. 649-654.

Christiansen, E.A., 1968, Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 5, p. 1167-1173,

Christiansen, E.A. 1979, The Wisconsinan glaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Christiansen, E.A. 1992, Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada: an update: Canadian Journal of Earth Sciences, v. 29, p. 1767-1778, Christiansen, E.A., and Sauer, E.K., 1989, Age of the Frenchman Valley and associated drift south of the Cypress Hills, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 25, p. 1703-1708.

Clayton, Lee, 1964, Karst topography on stagnant glaciers: Journal of Glaciology, v. 4, p. 107-112.

Clayton, Lee, 1967. Stagnant-glacier features of the Missouri Coteau in North Dakota. *in* Clayton, Lee, and Freers, T.F., eds., Glacial Geology of the Missouri Coteau and adjacent areas: North Dakota Geological Survey miscellaneous Series 30, p. 25-46.

Clayton, Lee, and Moran, S.R., 1974, A glacial process form model. *in* Coates, D.R., ed, Glacial Geomorphology, State University of New York, Binghampton, p. 88-119.

Clayton, Lee, and Moran, S.R., 1982, Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55-82,

Dowdeswell, J.A., and Sharp, M.J., 1986, Characterization of pebble fabrics in modern terrestrial glacigenic sediments: Sedimentology. v. 33, p. 699-710.

Dreimanis, A.E., 1989, Tills: their genetic terminology and classification. *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits: A.A. Balkema, Rotterdam, p. 17-84.

Driscoll, F.G., 1980, Wastage of the Klutan ice-cored moraines, Yukon Territory, Canada: Quaternary Research, v. 14, p. 31-50.

Dyke, A.S., and Prest, V. K., 1987, Late Wisconsinan and Holocene history of the Laurentide ice sheet: Geographie Physique et Quaternaire, v. 41, p. 237-263.

Elson, J.A., 1961, The geology of tills: *in* Penner, E. and Butler, J., eds., Proceedings of the 14th Canadian Soil Mechanics Conference, National Research Council of Canada, Associate Committee on Soil and Snow Mechanics Technical Memorandum 69, p. 5-36.

Elson, J.A., 1989, Comment on glaciotectonite, deformation till and comminution till: *in* Goldthwait, R.P. , and Matsch, C.L., eds., Genetic classification of glacigenic deposits: A.A. Balkema, Rotterdam, p. 85-88.

Evenson, E.B., Dreimanis, A., and Newsome, J.W., 1977, Subaquatic flow till: a new interpretation of the genesis of some laminated till deposits: Boreas, v. 6, p. 115-130.

Eyles, N., 1979, Facies of supraglacial sedimentation on Icelandic and Alpine temperate glaciers: Canadian Journal of Earth Sciences, v. 16, p. 1341-1361.

Eyles, N., 1983, Modern Icelandic glaciers as models for "hummocky moraine" in the Scottish Highlands. *in*: Evenson, E.D., Schluchter, Ch., and Rabassa, J., eds., Tills and related deposits: A.A. Balkema, Rotterdam, p. 47-60.

Eyles, N., Sladen., J., and Gilroy, S., 1982, A depositional model for stratigraphic complexes and facies superimposition in lodgement tills: Boreas, v. 11, p. 317-333.

Fenton, M.M., 1984, Quaternary stratigraphy of the Canadian Prairies, *in*, Fulton, R.J., ed., Quaternary Stratigraphy of Canada - a Canadian Contribution to IGCP Project 24. Geological Survey of Canada, Paper 84-10, p. 57-68,

Fullerton, D.S., and Colton, R.B., 1986, Stratigraphy and correlation of the glacial deposits on the Montana Plains. *in* G.M. Richmond and D.S. Fullerton, eds., Quaternary Glaciations of the United States: Quaternary Science Reviews, v. 5, p. 69-82.

Furnival, G.M., 1948, Cypress Lake map-area, Cypress Lake, Saskatchewan. Geological Survey of Canada, Memoir 242, 161p.

Ghibaudo, G., 1992, Subaqueous sediment gravity flow deposits: practical criteria for their field description and classification: Sedimentology, v. 39, p. 423-454.

Gibbard, P., 1980, The origin of the Catfish Creek Till by basal melting: Boreas, v. 9, p. 71-85.

Gillberg, G., 1977, Redeposition: a process in till formation: Geologiska Föreningens in Stockholm Förhandlingar, v. 99, p. 246-253.

Goldthwait, R.P., 1968, Introduction to till, today. *in* Goldthwait. R.P. , ed., Till: a symposium: Ohio State University Press. p. 3-27.

Gravenor, C.P., and Kupsch, W.O., 1959, Ice disintegration features in western Canada: Journal of Geology, v. 67, p. 48-64.

Haldorsen, S., and Shaw, J., 1982, The problems of recognizing meltout tills: Boreas, v. 5, p. 261-267.

Hartshorn, J.H., 1958, Flow till in southern Massachusetts: Bulletin of the Geological Society of America, v. 69, p. 477-482,

Hubert, J.F., and Filipov, A.J., 1989, Debris flow deposits in alluvial fans on the west flank of the White Mountains, Owens Valley, Calif., U.S.A: Sedimentary Geology, v. 61. p. 177-205.

Johnston, W.A. and Wickenden, R.T., 1931, Moraines and glacial lakes in southern Saskatchewan and southern Alberta, Canada: Transactions of the Royal Society of Canada, Section 4: Geological Sciences including Mineralogy, v. 69, p. 29-44.

Johnson, W.A., Wickenden, R.T., and Weir, J.D., 1948, Surface deposits of southwestern Saskatchewan: Geological Survey of Canada, Paper 48-18.

Klassen, R.W. 1984, Quaternary geology of southwestern Saskatchewan: *in* Current Research, Part A, Geological Survey of Canada. Paper 84-1A, p. 641-642.

Klassen, R.W., 1991, Surficial geology and drift thickness, Cypress Lake (72F), Saskatchewan: Geological Survey of Canada, Map 1766A (1:250,000 scale).

Klassen, R.W., 1992, Nature, origin, and age relationships of landscape complexes in southwestern Saskatchewan: Geographie Physique et Quaternaire, v. 46, p. 361-388,

Klassen, R.W., and Vreeken, W.J., 1985, Quaternary Geology of southwestern Saskatchewan: *in* Current Research, Part A, Geological Survey of Canada, Paper 85-1A, p. 187-192,

Klassen, R.W. and Vreeken, W.J. 1987, The nature and implications of surface tills and post-till sediments in the Cypress Lake area, Saskatchewan: *in* Current Research, Part A, Geological Survey of Canada, Paper 87-A, p. 111-125.

Kulig, J.J., 1985, A sedimentation model for the deposition of glacigenic deposits in central Alberta. [M.S. thesis], University of Alberta, 245 p.

Lawson, D.E., 1979a, A sedimentological analysis of the western terminus of the Matanuska Glacier, Alaska: U.S. Army Cold Regions Research and Engineering Laboratory. Report 79-9. 122pp.

Lawson, D.E., 1979b, A comparison of the pebble orientations in ice and deposits of the Matanuska Glacier, Alaska: Journal of Geology, v. 87, p. 629-645.

Lawson, D.E., 1981, Distinguishing characteristics of diamictons at the margins of the Matanuska Glacier, Alaska: Annals of Glaciology, v. 2, p. 78-84,

Lawson, D.E., 1989, Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits, *in* Goldthwait, R.P., and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam, p. 147-170.

MacKenzie, G.D., 1969, Observations on the collapsing kame terrace in Glacier Bay National Monument, south eastern Alaska: Journal of Glaciology, v. 2, p. 413-425.

McDonald, B.C., and Shilts, W.W., 1975, Interpretation of faults in glaciofluvial sediments, *in* Jopling, A.V. , and MacDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation, Tulsa OK, S.E.P.M. Special Publication 23, p. 123-132.

Marcussen, I., 1975, Distinguishing between lodgement and flow till in Weischselian deposits: Boreas, v. 4, p. 113-123.

Mark, D.M., 1973, Analysis of axial orientation data, including till fabrics: Geological Society of America Bulletin, v. 84, p. 1369-1374,

Mark, D.M., 1974, On the interpretation of till fabrics: Geology, v. 8, p. 101-104.

May, R.W., 1977, Facies model for sedimentation in the glaciolacustrine environment: Boreas, v. 6, p. 175-180,

McConnell, R., 1885, Report on the Cypress Hills, Wood Mountain, and adjacent country: Geological Survey of Canada, Annual Report, 1C.

Mickelson, D.M., 1973, Nature and rate of basal till deposition in a stagnating ice mass, Burroughs Glacier, Alaska: Arctic and Alpine Research, v. 5, p. 17-27.

Moran, S.R., 1971. Glaciotectonic structures in drift, *in* Goldthwait, R.P., ed., Till: a Symposium. Ohio State University Press, p. 127-148.

Odynsky. W., and Newton, J.D., 1950, Soil Survey of the Wycroft and Watino sheets: Alberta Soil Survey Report 15, 84p.

Odynsky, W., Wynn, K.A., and Newton, J.D., 1952, Soil survey of the High Prairie and McChennan Sheets: Alberta Soil Survey Report 17, 112p.

Ostrem, G., 1959, Ice melting under a thin layer of moraine and the existence of ice cores in moraine ridges: Geografiska Annaler, v. 41, p. 228-230.

Ovenshine, A.T., 1970, Observations of iceberg rafting in Glacier Bay, Alaska and identification of ice-rafted deposits: Geological Society of America Bulletin, v. 81, p. 891-894.

Paul, M.A., and Eyles, N., 1990, Constraints on the preservation of diamict facies (melt-out tills) at the margins of stagnant glaciers: Quaternary Science Reviews, v. 9, p. 51-69.

Proudfoot, D.N., 1985, A lithostratigraphic and genetic study of Quaternary sediments in the vicinity of Medicine Hat, Alberta. [Ph.D. thesis], Edmonton, Alberta, University of Alberta, 345 p.

Rappol, M., 1985, Clast-fabric strength in tills and debris flows compared for different environments: Geologie en Mijnbouw, v. 64, p. 327-332.

Ronnert, L., 1992, Genesis of diamicton in the Oak Creek Formation of south-east Wisconsin, U.S.A.: Sedimentology, v. 39, p. 177-192,

Rusczyñska-Szenajach, H., 1983, Lodgement tills and syndepositional glaciotectonic processes related to subglacial thermal and hydrologic conditions, *in*: Evenson, E.D., Schluchter, Ch., and Rabassa, J., eds., Tills and related deposits, A.A. Balkema, Rotterdam, p. 113-117.

Rusczyñska-Szenajach, H., 1987, The origin of glacial rafts: detachment, transport, deposition: Boreas, v. 16, p. 101-112.

Rusczyñska-Szenajach, H., 1988, Glaciotectonics and its relationship to other glaciogenic processes: *in* Croot, D.G., ed., Glaciotectonics: forms and processes, A.A. Balkema, Rotterdam. p. 191-193,

Rust, B., R., and Romanelli, R., 1975, Late Quaternary subaqueous outwash deposits near Ottawa, Canada. *in* Jopling, A.V., and MacDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation, Tulsa OK, S.E.P.M. Special Publication 23, p. 177-192.

St.Onge, D.A., 1972, Sequence of glacial lakes in north-central Alberta: Geological Survey of Canada Bulletin 213 16 pp.

Sharp, R.P. , 1949, Studies of supraglacial debris on valley glaciers: American Journal of Science, v. 247, p. 289-315.

Shaw, J., 1982, Meltout till in the Edmonton area, Alberta, Canada: Canadian Journal of Earth Science, v. 9, p. 1548-1569.

Shaw, J., 1983, Forms associated with boulders in melt-out till,. *in* Evenson, E.B., Schlütcher, Ch., and Rabassa, J., eds., Tills and Related Deposits. A.A. Balkema, Rotterdam, p. 3-13.

Shaw, J., 1987, Glacial sedimentary processes and environmental reconstruction based on lithofacies: Sedimentology, v. 34, p. 103-116.

Shetsen, I. 1984, Application of till pebble lithology to the differentiation of glacial lobes in southern Alberta: Canadian Journal of Earth Sciences, v. 21, p. 920-933.

Stalker, A. MacS., 1960, Surficial geology of the Red Deer-Stetler map area: Geological Survey of Canada Memoir 306, 140p.

Stalker, A. MacS., 1973, Surficial geology of the Drumheller area, Alberta: Geological Survey of Canada Memoir 370, 122 p.

Stalker, A.MacS. 1977, The probable extent of classical Wisconsinan ice in southern and central Alberta: Canadian Journal of Earth Sciences, v. 14, p. 2614-2619.

Stalker, A.MacS., and Harrison, J.E., 1977, Quaternary glaciation of the Waterton-Castle River region of Alberta: Canadian Petroleum Geology Bulletin, v. 25, p. 882-906.

Shetsen, I. 1984, Application of till pebble lithology to the differentiation of glacial lobes in southern Alberta: Canadian Journal of Earth Sciences, v. 21, p. 920-933.

Stone, B.F., 1976, Analysis of slump slip lines and deformation fabric in slumped Pleistocene lake beds: Journal of Sedimentary Petrology, v. 46, p. 313-325.

Visser, J.N.J., Collistan, W.P. , Terblanche, J.C., 1984, The origin of soft sediment deformation structures in Permo-Carboniferous glacial and pro-glacial beds, South Africa: Journal of Sedimentology, v. 54, p. 1183-1196.

Vreeken, W.J. 1986, Quaternary events in the Elkwater area of southeastern Alberta: Canadian Journal of Earth Sciences, v. 23, p. 2024-2038.

Watson, R.A., 1980, Landform development on moraines of the Klutan Glacier, Yukon Territory, Canada: Quaternary Research, v. 14, p. 50-60.

Weertman, J., 1963, Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets: Journal of Glaciology, v. 25, p. 457-476,

Westgate, J.A., 1965a, The surficial geology of the Cypress Hills area, Alberta: Alberta Research Council Preliminary Report 65-2, 6p.

Westgate, J.A., 1965b, The Pleistocene stratigraphy of the Foremost-Cypress Hills area Alberta: Alberta Society of Petroleum Geologists, 15th Annual Field Conference, Guidebook, Part 1, Cypress Hills Plateau, p. 85-111.

Westgate, J.A., 1968, Surficial geology of the Foremost - Cypress Hills area, Alberta: Research Council of Alberta Bulletin 22, 122 pp.

Westgate, J.A., 1972, The Cypress Hills. *in* Quaternary geology and Geomorphology between Winnipeg and the Rocky Mountains. International Geological Congress, 24th Session Montreal Quebec, Guidebook for Excursion C-22, p. 50-62.

Whitaker, S.H. 1965, Geology of the Wood Mountain Area (72G) Saskatchewan. [Ph.D. Thesis], Urbana, IL, University of Illinois.

Whitaker, S.H., 1976, Geology and groundwater resources of the Cypress area (72-F), Saskatchewan: Saskatchewan Research Council Geology Division, Map 22 (1:250,000 scale).

Chapter 3. Glaciolacustrine diamicton on and around the south flanks of the Cypress Hills and its relevance to the deglaciation chronology of the area.

A version of this paper will be submitted for publication in the Canadian Journal of Earth Sciences.

Introduction

The glacial deposition processes that were active on and around the southern flanks of the Cypress Hills in Saskatchewan can be investigated by examining exposures of glacigenic diamicton and normally graded silt beds there. The diamicton beds range from apparently unbedded to stratified units in which thin sub-horizontal diamicton beds (5 to 20 cm thick) are intercalated with normally-graded silt and clay laminae. Glaciolacustrine rhythmites are associated with these units. The objectives of this study are to determine the origin of the diamicton beds and to determine their relationship to the rhythmites. The importance of these deposits to the interpretation of local glacial events, such as the deposition of erratics beyond the Late Wisconsinan margin and formation of the Frenchman channel are discussed. The possibility of observing similar stratified glacigenic diamicton units throughout the Prairies is also examined.

Location, geologic setting, and history

The Cypress Hills is a preglacial plateau rising up to 430 m above the prairie surface (maximum elevation 1410 m). They consist of three blocks: the West Block, which extends 70 km into Alberta, and the Center and the East blocks, located entirely in southwest Saskatchewan (Fig. 3.1A). The West and Center blocks have steep northern margins and more gently sloping southern flanks, whereas the East Block on its eastern and northern flanks, rises gently from the prairie surface.

Prior to Klassen (1991, 1992) and this study, the glacial history of the Cypress Lake area had been largely inferred from studies of peripheral areas. These local and regional stratigraphic studies (Westgate 1965a,b, 1968, 1972; Christiansen 1979; Clayton and Moran 1982; Fullerton and Colton 1986; Dyke and Prest 1987a, b) show that the Late Wisconsinan ice extended through the study area and terminated in Montana. Klassen (1992) provided a more accurate depiction of ice-lobe distribution, indicating that at the Late Wisconsinan maximum, the Cypress Hills were surrounded by three ice lobes (West, Gap, and East lobes, Fig. 3.1A). Ice reached a maximum elevation of 1350 m on the north side of the Cypress Hills in Alberta (Westgate 1968). The upper 100 to 200 m of the hills projected above the ice as nunataks. Scattered glacigenic deposits and glacial landforms from this ice mass are found across the southern flanks, up to about 1190 m elevation (Klassen 1992 and this study). Erratics are present up to 1250 m elevation on the south side of the West Block. Klassen (1992) described six landscape complexes in the area, reflecting different processes and times of formation.

Westgate (1968) identified two drift sheets on the southern flanks of the Cypress Hills of Alberta to the west. These sheets should have equivalents in this study area. The Elkwater drift, a patchy deposit interpreted as the remnants of a pre-late Wisconsinan ice sheet, probably corresponds to the "residual drift landscape complex" of Klassen (1992). The Wildhorse drift (Westgate 1968), a morphostratigraphic unit, was deposited around the Late Wisconsinan maximum. It may correspond to the "last advance landscape complex" of Klassen (1992). The deposits of the Late Wisconsinan ice in the study area should be temporally equivalent to the Late Wisconsinan Fort Assiniboine and Loring tills of northern Montana (Fullerton and Colton 1986). These tills are considered by Fullerton and Colton (1986) to be equivalent to the Late Wisconsinan Wymark till of southern Saskatchewan (Christiansen 1968). Formal lithostratigraphic names and type sections have not yet been designated for the tills in the study area.

The Frenchman channel, the most prominent glacial landform in the area, extends across slope along the southern flanks of the Cypress Hills and Wood Mountain uplands and terminates at the Milk River in Montana (Fig. 3.1A). Christiansen (1979), Clayton and Moran (1982), and Dyke and Prest (1987a) depicted the Frenchman channel as an interlobate channel, separating ice that flowed around the hills from the east, from ice that flowed around the hills from the west. Their depiction showed that the channel was gradually incised by the meltwater it carried. Klassen (1992) placed a more extensive East Lobe, which terminated in Montana, over the Frenchman channel. This ice extended onto the south flanks of the East and Center blocks. Since the channel area was ice covered, it was not an interlobate feature. It has been reinterpreted as a spillway, cut by catastrophic outbursts from ice-dammed lakes impounded on and near the Cypress Hills (Chapter 4, this volume).

Fieldwork

Three sections were investigated for this study (Fig. 3.1A): the Canal section at 1100 m elevation along the western edge of the East Block. The UTM grid co-ordinates for the Canal section are 72F/11,12UXK178918 for the south end of the section and 177929 for the northern terminus of the section (LSD W of 3, T7 R25 S19,1/4SEC. SE quadrant of 8 to SE quadrant of 16). The Gilchrist section at 1050 m elevation on the east side of the Center Block (UTM co-ordinates 72F/11,12UXK166851) (LSD: W of 3, T6 R26 S36 1/4 sec. SW quad. of 4), and the Cypress Lake section, composed of several smaller exposures located along the southwest edge of Cypress Lake south of the Center Block (UTM co-ordinates 72F/6,12UXK178918 to 177929) (LSD W of 3, T6 R27 S19 1/4 sec SEquad. of 6 to NW quad. of 7) at 990 m elevation. Field traverses and air photograph analysis were used to trace the glacigenic sediment beyond the sections and to identify landforms such as moraines, meltwater channels, terraces, and ice pushed ridges. Glaciofluvial sediment exposed in gravel pits was described and photographed. The elevational limit of granite, gneiss, and carbonate erratics on the southern flank of the West Block in the Graburn Gap area was also determined.

Sedimentary units in the sections were described, photographed, and sampled. Fabric analysis was performed on diamicton units by measuring the a-axis orientation of a minimum of 25 phenoclasts with a/b axial ratios of at least 1.5:1 and minimum a-axis lengths of 2 cm. The results were plotted as unidirectional rose diagrams on Schmidt nets and interpreted using the techniques described in Mark (1973, 1974).

Impounded glacial lakes.

Glacial Lake Belanger

Glacial Lake Belanger (hereafter called Lake Belanger) formed in the depression between the East and Center blocks (Fig. 3.1B) when the East Lobe advanced upslope onto the south flanks of the Cypress Hills. This ice impounded meltwater from the ice mass north of the Cypress Hills. The lake is named for Belanger Creek, which flows southward through the area occupied by the former lake. A composite section containing three units (Figure 3.2) was constructed. The Gilchrist Ranch section (Fig 3.1A), an east-facing exposure 30 m long and 7 m high on Sucker Creek, contains Units 1 and 2, which unconformably overlie the Cypress Hills Formation described in Von Hoff (1969) and Leckie and Cheel (1989). The Canal section along the west edge of the East Block is over 900 m long and contains from 4 to 6 m of glacigenic sediment in two units (Units 2 and 3). These units also unconformably overlie the Cypress Hills Formation.

Unit 1. Normally-graded silt beds with clay interlayers.

Unit 1 is about 1.2 m thick, extends about 20 m across the exposure, and contains numerous normally-graded silt laminae, clay beds, and couplets of silt and clay. Sharp contacts separate sub-horizontal silt laminae that range from 1 mm to 8 mm thick. These laminae form beds 10 to 30 cm thick that extend the length of the exposure (20 m). Silt beds (composed of multiple laminae) are commonly separated by 1 to 3 cm thick clay beds (Fig. 3.3, 3.4). The texture of the clay beds was 1.9% sand, 26% silt, and 72% clay. Silt beds averaged 4% sand, 68% silt, and 28% clay (see Appendix 3). The contacts between clay beds and overlying silt laminae are sharp whereas the contacts between the silt laminae and overlying clay beds are commonly gradational. Phenoclasts make up 2 to 3% of the volume of Unit 1. Only 1 to 2% of all phenoclasts are erratics. Phenoclasts often deform the underlying laminae and silt and clay couplets. Fifty to one hundred centimetres below the contact with Unit 2, the clay and silt layers are broken and slickensided (Fig. 3.4). The 20 to 30 cm long disruption zones plunge 10^0 to 15^0 to the south.

Genesis of sediment in Unit 1.

The silt laminae observed in Unit 1 are similar to graded silt layers described from lacustrine environments (Lambert and Hsu 1979). The lake must have been a glacial lake as shown by the sidehill location of the deposit and the presence of glacial erratics. The normal grading in the silt laminae and the sharp contacts between them match similar sediments interpreted as suspension deposits derived from the tails of turbid undercurrents deposited in ice-proximal lakes (Ashley 1975, 1989, Banerjee 1973). The sharp contacts between successive silt laminae indicate erosive planation by the successive undercurrents or pulsating flow in a single undercurrent (Smith and Ashley 1985). The clay laminae were deposited from suspension from either the tails of these underflows or through the water column. The silt-clay couplets reflect deposition from the tail of the turbidity current or a combination of turbidity current deposition (normally-graded silt layers) with suspension deposition through the water column, of all or part, of the clay layer (Smith and Ashley 1985). The presence of several silt laminae between each silt and clay couplet indicates deposition from several undercurrents or from a pulsing undercurrent. Some of the normally-graded silts may have been deposited from slump-generated underflows (Hampton 1972; Smith and Ashley 1985). It is difficult to separate slump-generated underflow rhythmites from rhythmites deposited directly from fluvial underflow currents (Smith and Ashley 1985). No attempt was made in this study to separate suspension sediments deposited from the tail of a turbidity current from those that have settled through the water column. Ashley (1989) reported that laminated, normally-graded silt laminae and rhythmites were deposited in the intermediate and distal parts of lake basins and stated that "sediments transported and deposited by underflow are likely to contain current bedding structures in proximal areas and to consist of multiple graded beds in the distal portions of the lake" (Ashley 1989, p. 257). The presence of multiple normally-graded silt laminae without bedding structures in Unit 2 favors a distal depositional setting.

Lambert and Hsu (1979) noted that underflows can occur sporadically throughout the year depositing more than one layer of graded sediment in a single year. Gilbert and Shaw (1981) observed diurnal deposition of normally-graded silt laminae in a pro-glacial lake. The exact time scale for deposition of the laminae and couplets in Unit 1 therefore cannot be determined. It is therefore incorrect to classify them as varves which are silt and clay couplets deposited over a one year cycle.

The large phenoclasts with underlying deformation structures are dropstones that were ice rafted into the lake. The disruptions of the clay and silt laminae (Fig. 3.3, 3.4) trend upslope, therefore they are not shear planes caused by gravitational movements and more likely reflect overriding of the deposit by ice flowing upslope from the south. It is possible that deposition of the upper parts of Unit 1 or deposition of Unit 2 could have initiated shearing in the basal zone of Unit 1.

Unit 2. Unbedded stony diamicton

At the Gilchrist Ranch section, Unit 2 is a 2 to 3 m thick, oxidized olive yellow (2.56/6 dry) clayey diamicton. The exposures length is 8 m at the base and 3 m at its top. A 30 cm thick gradational sequence is observed between Units 1 and 2 at Gilchrist Ranch. The sequence contains thin, discontinuous, sub-horizontal diamicton beds of (up to 8 cm thick and extending 2 to 3 m in section) interlayered with normally-graded silt laminae. In the upper 10 cm of the sequence, the diamicton beds are separated by a single silt lamina (Fig. 3.5). In the Canal section, Unit 2 lies in a trough in the underlying Cypress Hills Formation.

The unit consists of an unoxidized black (2.5Y 2.5/1 dry) unbedded well-consolidated diamicton more than 3.5 m thick that extends over 5 m in section. The average grain size distribution of the fraction finer than pebble-size is 34% clay, 33% silt and 31% sand at the Canal section and 36% clay, 36% silt, and 28% sand at the Gilchrist section (see Apendix 3). The diamicton contains 10 to 15% phenoclasts by volume. Of these, 3 to 4 % are granite and gneiss transported from the Canadian Shield or carbonate erratics from the edge of the shield. The contact between Units 2 and 3 was not exposed in the Canal section. In both sections the diamicton is dense and structureless and lacks shale pellets.

Thin-sections from Unit 2 reveal numerous rounded and sub-rounded diamicton intraclasts (these intraclasts have the same textural make-up as the surrounding diamicton, but lack sediment coarser than sand, have a matrix (fine silt and clay) percentage of more than 60%, and do not show alignment of plasma around phenoclasts (insepic plasma fabric, Brewer 1976)). Shale phenoclasts are absent. The lower 1 m of Unit 2 at Gilchrist Ranch, contains numerous plate and blade-shaped phenoclasts. The a/b axial planes of these phenoclasts are predominantly horizontal (160 of 170 phenoclasts counted had horizontal a-b axial planes, the a/b axial planes of the remaining 10 phenoclasts were less than 5^{0} from horizontal, Fig. 3.5). Phenoclasts are commonly clustered in small groups. No disruption of laminae was observed beneath the phenoclasts. Fabrics from Unit 2 at the Canal section have low S_1 eigenvalues and do not indicate preferred orientation (Table 3.1).

The genesis of the sediments in Unit 2

Gradational Zone

The paucity of sedimentary structures makes genetic interpretation of Unit 2 difficult. Walker (1984) stated that gradational contacts between sedimentation units indicate a gradual shift in sedimentary processes and depositional environment. The gradational contact between Unit 1 (graded silt and clay laminae) and Unit 2 (unbedded stony diamicton) can be interpreted to indicate a gradual shift from a distal basinal glaciolacustrine environment to a more proximal one. In the distal environment, suspension deposition predominated (deposition from underflows in distal positions was dominated by suspension deposition from the tail of the underflow, suspension deposition through the water column, the two types of suspension deposits were not separated in this study). In proximal areas, deposition was dominated by the emplacement of glaciolacustrine diamicton layers and proximal underflow sediments. This transition reflects ice advance towards the area. The diamicton beds of Unit 2 were deposited in a glacial lake either by debris flows (Evenson et al. 1977), by rainout beneath an undermelted ice margin (Gibbard 1980), or a combination of the two (Dreimanis 1982). These diamictons are referred to here as glaciolacustrine diamictons to differentiate them from glacigenic debris-flow diamictons deposited subaerially. Calling the diamicton layers subaquatic flow tills would obscure the fact that the sediments were deposited in a glaciolacustrine environment. The thin diamicton beds with parallel planar contacts are similar to fluid debris-flow deposits described by Bull (1964) and Lawson (1979a, 1989). The diamicton intraclasts in the diamicton are often considered diagnostic of debris flows (Eyles 1979; Kulig 1985) but can also be found in ice-rafted or rain-out deposits (Ovenshine 1970). The sub-horizontal phenoclast orientation in the lower 1 m of Unit 2 at the Gilchrist section is not diagnostic of glaciolacustrine deposition. Horizontal phenoclasts fabrics have been reported from fluid, subaerial debris-flows (Bull 1964; Lawson 1979b), from meltout till (Boulton 1975; Shaw 1979, 1982) as well as for dropstones (Schmoll 1961; Griggs and Kulm 1969). The size and shape of the phenoclasts preclude the likelihood that they rolled into position. The deformation of underlying graded silt laminae makes meltout and subaerial debris-flow deposition unlikely. Very fluid, laminar debris flows commonly have few phenoclasts and uniform sandy clay textures (Bull 1964; Pierson 1980, 1981; Lawson 1979b, 1981). Additionally fluid debris-flows are not competent to carry large phenoclasts (Pierson 1980, 1981) such as those in Unit 2 The horizontal phenoclast orientation favors a dropstone origin as does clustering of the phenoclasts, considered by Ovenshine (1970) and Thomas and Connel (1985) to be characteristic of iceberg-rafted sediment

The final resting attitude of dropstones is influenced by phenoclast shape and the nature of the substrate. Rod and ovoid-shaped phenoclasts descend through the water column with their axis horizontal, while discoid and platy dropstones follow unstable fall paths (Stringham et al. 1969, Domack and Lawson 1985). Dropstones, with vertical or high impact angles, become embedded if the substrate is soft, and topple onto their sides if it is consolidated. The lack of dropstone impact structures beneath phenoclasts indicates that the substrate was consolidated (Domack and Lawson 1985). The phenoclasts subsequently toppled onto their sides. The lower part of Unit 2 is, therefore, an assemblage consisting of thin fluid debris-flow deposits, ice-rafted and rain-out sediment interlayered with silt laminae deposited by suspension from the tails of turbidity currents.

Unbedded Zone

The unbedded, upper part of Unit 2 in the Canal and Gilchrist sections is difficult to interpret. The gradational contact between Units 1 and 2 favors a gradual shift in sedimentary process within similar sedimentary environments (Walker 1984, p. 6). Without a dramatic shift in process or environment, the massive upper part of Unit 2 should reflect proximal deposition. Proximal deposition is characterized by an increase in debris-flow sedimentation (Evenson et al. 1976). Since diamicton intraclasts have been associated with both debris-flows (Lawson 1979) and rain-out deposition (Ovenshine 1970) it is unclear which process is dominant. The phenoclast fabric strengths are similar to those of debris-flows (Dowdeswell and Sharp 1986; Lawson 1989; Evenson et al. 1976) or rain-out sediment (Gibbard 1980). If the unit is composed of coalesced debris-flows, an unbedded deposit could be formed if there was insufficient time between debris flow events to deposit a recognizable thickness of contrasting silt and clay interbeds or through the emplacement of a single large debris flow. The absence of intervening silt and clay laminae could also indicate a decrease in underflow activity or a shift in depositional area of the underflows. It is also possible that thin underflow deposits were eroded and incorporated into the debris flows as they flowed over them. Formation of an unbedded glaciolacustrine rain-out diamicton (Gibbard 1980) could occur if there was a rapid input of a large volume of debris without intervening turbid underflows.

Unit 3 Diamicton beds intercalated with clay and normally-graded silt laminae

Unit 3, exposed throughout the Canal Section, is composed of diamicton beds intercalated with beds of normally-graded silt and laminae and beds of clay (Fig. 3.2, 3.6, 3.7). The unit is 3 to 4 m thick and overlies either the Cypress Hills Formation or Unit 2 (unbedded stony diamicton). Near the section top, normally-graded silt beds predominate.

The diamicton beds have a variety of geometries ranging from horizontal beds 10 to 20 cm thick and extending over 10 m in section, to plano-convex pods up to 1 m thick and extending for 5 m. The diamicton beds near the base of the unit are dark grayish brown (2.5Y4/2 dry), structureless, contain 15 to 20% phenoclasts by volume and possess an average of 26% clay, 47% silt, and 31% sand. Diamicton beds near the base of the unit contain diamicton intraclasts. Two samples from silty diamicton beds near the top of the unit, (from near the northern end of the canal), are light yellowish brown (2.5Y6/4), contain only 1 to 2% phenoclasts, and are primarily composed of irregularly-shaped, angular silt and clay intraclasts within a silt matrix. The average texture of these samples is 11% sand, 43% silt,and 46% clay (see Appendix 3). Shale phenoclasts are very rare in the diamicton beds and in thin sections from these diamictons.

Diamicton layers are commonly separated by clay layers (Fig. 3.7, 3.8). Clay layers in the lower part of Unit 2 consist of either a single clay lamina or several thin clay laminae. They are commonly discontinuous, extending laterally for a few tens of centimeters to more than 20 m (Fig. 3.7, 3.8). Clay beds 2 to 8 cm thick, in the upper 1.5 to 2 m of the unit, are formed of thin clay laminae. Thin, normally-graded silt beds are distributed randomly within the clay beds (Fig. 3.9). Because the silt and clay layers conformably drape diamicton beds, they have a wide range of attitudes that reflect the irregular upper surfaces of the underlying diamicton beds. A sub-horizontal clay bed directly overlying the Cypress Hills Formation maintained its sub-horizontal attitude as the top of the Cypress Hills Formation dropped. A diamicton layer occupied the space between the clay and the preglacial gravel. Near the top of Unit 3 the clay and silt beds are commonly more horizontal and continuous. Disruption and contortion of some of these upper silt and clay beds, causing overthickening, was observed.

The contacts between diamicton beds and underlying clay beds are commonly irregular with parts of the diamicton layer extending downward into the clay. Large phenoclasts (greater than 5 cm along the a-axis) commonly deform both the clay beds and the diamicton beds immediately underlying them. The deformation mirrors the shape of the phenoclast (Fig. 3.10). Phenoclast fabrics measured in Unit 3 have low S_1 eigenvalues, are multimodal, and do not show a preferred orientation (Table 3.1).

Genesis of the sediments in Unit 3

Determining the genesis of the intercalated stratified layers permits a better understanding of the formation of the diamicton unit. The normal grading and continuity of the silt beds that drape the diamicton beds regardless of the attitude of the underlying surface, indicate that the silt layers were deposited from suspension. These suspension sediments could have been deposited by settling through a water column or by suspension from the tail of a turbid underflow (Ashley 1975, 1989; Banerjee 1973). Silt laminae with sharp basal contacts were probably deposited from turbid underflows. Clay laminae, that also drape diamicton beds, preglacial gravels, and graded-silt laminae, were deposited between debris-flow events by suspension from the tails of turbid underflows or through the water column. The large cobbles and boulders that deform the clay and silt interlayers are dropstones. The irregular geometry of the diamicton beds, their irregular upper surfaces and the rapid change in their thicknesses and attitudes over short distances match descriptions of muddy debris-flow deposits (Ghibaudo 1992). Their inconsistent, weak phenoclast fabrics also correspond to those expected for debris-flow deposits (Dowdeswell and Sharp 1986; Lawson 1979a, b, 1989). The intercalation of subaqueous silt and clay laminae with diamicton layers indicates that the debris-flows were deposited subaqueously. Episodic deposition of the debris-flows allowed clay suspension layers to drape the upper surfaces of the debris-flows. The multiple attitudes of the clay beds also indicate suspension deposition over irregular surfaces. Loading features in the clay interlayers beneath diamicton beds indicates that the clay layers were only partially consolidated when the diamicton bed was deposited on them. Though the sediments of Unit 3 are glacially derived, they have been modified by glaciolacustrine processes, it is best to consider the unit a subaqueous glacigenic debris-flow complex deposited in an ice-contact lake. This terminology provides a complete picture of sedimentation and the sedimentary environment whereas subaquatic flow till tells little about the depositional processes that formed the deposit and stresses a glacial origin over a glaciolacustrine genesis.

The upsection increase in thickness of clay and graded silt beds suggests a gradual reduction in debris-flow input and increasing deposition from suspension and turbid underflows. Diamicton beds composed of silt and clay intraclasts in a silt matrix near the top of the unit indicate that the normally-graded silts and clay layers were unstable and prone to failure. Liverman (1981) observed that the unconsolidated clay and silt within an ice-dammed lake were subject to slumping and multiple rotational failures when the lake drained. The silt and clay intraclast diamicton layers could have formed when Lake Belanger drained. Multiple drainage events have been reported for ice-dammed lakes (Howarth 1968; Gilbert 1971; Lindsay 1966; Stone 1963). Because there are several such beds near the top of Unit 3, the lake may have undergone several drainage events.

The gradational sequence between Units 2 and 3 indicates that they are interrelated (Walker 1984) and that no major shift in depositional environment or processes occurred. This indirectly supports an interpretation that Unit 2 is a coalesced debris-flow deposit. An increased time interval between debris-flow events or a slight change in process or loci of deposition of the underflows would transform the unbedded diamicton of unit 2 into the stratified diamicton complex of Unit 3. The superimposition of units in a proximal glaciolacustrine setting is discussed later.

Depositional summary

A gradual shift is observed from glaciolacustrine rhythmite and normally graded silt lamina deposition to deposition of the unbedded glaciolacustrine debris-flow diamicton of Unit 2. This reflects gradual encroachment of ice into this area of the lake. The normally-graded silt and clay of Unit 1 are interpreted to be more distal deposits. Over-riding by glacial ice sheared these underlying silt and clay deposits, creating the structures shown in Figure 3.4. It is possible that the unbedded portion of Unit 2 in places was deposited directly from the ice as a subglacial meltout till and that overriding of Unit 1 caused the shearing observed in Unit 1. The deformations could also have arisen through loading of the unstable saturated clays and silts when they were covered by a thick debris-flow originating from nearby glacier ice. At the ice front, underflow sediments and water column suspension sediment were intermixed with debris released, near or at the ice front, by iceberg calving, rainout or debris-flow. Gradually for a variety of reasons (short time interval between debris flow events, shifting depositional sites, thick proximal debris flows, or incorporation of underlying silt and clay laminae), sedimentation from rainout or by debris-flow predominated, resulting in deposition of the unbedded diamicton in the upper part of Unit 2. The reappearance of silt laminae and clay layers between debris-flow deposits (Unit 3) indicates a return to earlier sedimentation patterns. The appearance of silt and clay intraclast diamicton beds near the top of Unit 3 indicates a further shift to glaciolacustrine depositional processes. Suspension and underflow sedimentation were again dominant but these unstable stratified sediments were resedimented subaqueously possibly as a result of rapid lake drainage.

Extent of the Lake Belanger

The lacustrine deposits of the Canal Section are traceable in air photographs across the depression between the East and Center blocks. These deposits were superimposed upon small meltwater channels (0.5 to 2 m in depth and 5 to 10 m wide) that had transported meltwater across the Cypress Hills from the northern ice mass (Klassen 1991). The glaciolacustrine diamicton deposits can be traced westward to the vicinity of the Gap Lobe. There, they disappear and hummocky gravel containing few erratics are observed. The reason for this change from glaciolacustrine diamictons to hummocky gravels is uncertain. It may be that the thin (when compared to the ice covering the prairie surface), relatively debris-free ice bordering the lake in this area may have pushed and incorporated large volumes of preglacial gravels of the Cypress Hills Formation thereby diluting the volume of non-local basal material. Stagnation and meltout, then formed the ridged and hummocky gravel deposits that contain little matrix material and few erratics.

The northern limit of the lake is marked by a belt of hummocky terrain (Klassen 1991). As in the Gap area, the hummocky terrain closest to the lake margin is composed of ice-pushed, hummocky gravels with less than 0.1% erratics.

A meltwater channel, 100 m across and 5 to 10 m deep, is seen at 1125 m elevation on the western margin of the former lake (Fig. 3.1B). The channel is about 20 m above the floor of the former lake. It can be followed westward from the margin of the former lake (Fig. 3.1B) to a channel that carried meltwater along the eastern edge of the Gap Lobe. This wind gap channel permits the maximum elevation of at least one surface level of the former lake to be estimated. The channel is small and was not incised to the base of the former lake therefore it could not have completed drained Lake Belanger. There is no assurance that this was the highest drainage outlet for the lake. Complete drainage of the lake was through the Sucker Creek and Davis Creek coulees (Fig. 3.1B) which join the Frenchman channel east of Cypress Lake Reservoir. The lake was approximately 60 m deep, covered about 100 km^2, and impounded between $6.0X10^9$ and $1.0x10^{10}$ m^3 of water. Erratics in this area up to an elevation of 1125 m were probably ice-rafted.

Glacial Lake Cypress

Glacial Lake Cypress (hereafter called Lake Cypress) was located immediately south of the Center Block where the Frenchman channel broadens dramatically (Fig. 3.1B). It is named for Cypress Lake Reservoir which occupies part of the former lake's basin. The absence of the south margin of the Frenchman channel and impounding of Lake Cypress indicates that the south wall of the channel and south margin of the lake were formed by ice (Fig. 4.10). The Cypress Lake section is composed of several small exposures that overlie Cretaceous shale knobs. In the most extensive exposure, the glacigenic sediments vary from 2 to 8 m in thickness and extend for over 200 m along the southern margin of Cypress Lake Reservoir (Fig. 3.1A). There are two units. Unit 1 at the base is a diamicton containing abundant, soft friable shale and sandstone phenoclasts. Unit 1 is conformably overlain by Unit 2 which contains diamicton beds intercalated with normally-graded silt laminae and clay laminae. Smaller exposures to the west also contain these units but are less accessible and complete.

Cypress Lake section Unit 1: Diamicton with sand and gravel lenses and numerous shale phenoclasts

The diamicton of Unit 1 unconformably overlies Cretaceous shale and is 1 to 1.5 m thick on the flanks of the shale knoll and 0.2 to 0.4 m thick over its crest (Fig 3.11). Its lower contact is sharp and irregular. Since no contact with the overlying Unit 2 was discernible, the base of the lowest continuous silt laminae in Unit 2 was taken as the top of Unit 1.

The diamicton varies widely in grain size and compaction. In some areas, the diamicton is poorly compacted, stony (up to 40% phenoclasts), and has a sandy silt matrix. In others, it is dense and compact, has a clayey silt matrix and only 5 to 10 % phenoclasts. Where it is dense and compact it is very dark grayish brown (2.5Y3/2) and has 24 to 33% clay, 26 to 34% silt, 22 to 33% sand. Poorly sorted sand and sandy gravel lenses consisting of pebble-sized phenoclasts, disrupted discontinuous stringers, and pods of sand and gravel are present. Angular to subangular unstriated blocky shale phenoclasts derived from the underlying shale are common and locally make up 25 to 30% of the diamicton. Rare blocks of angular, non-local, friable, yellow, medium-grained sandstone are also present (Fig. 3.12). Erratics constitute less than 2% of all phenoclasts. The proportion of shale phenoclasts is highest near the base of the unit and decreases vertically with quartzite phenoclasts predominating near the top of the unit. Eigenvalues 0.46 and 0.53 with orientations of 235^0 and 345^0 respectively for the principal eigenvector (S_1) were recorded for two fabrics from Unit 1

Interpretation of genesis of the sediment in Unit 1

The origin of the sediments in Unit 1 is complex. Shale phenoclasts in Unit 1 derived from the underlying outcrop, represent a strong local lithological component at the base of the unit. The shale phenoclasts are angular and have not been sheared, smudged or deformed. Additionally, they are in close contact with non-local, angular friable sandstone cobbles. Both the shale and sandstone phenoclasts indicate that the unit was deposited by a non-disruptive process. Neither creep nor slopewash could have formed the dense unbedded diamicton that forms the bulk of Unit 1. Nor can creep or slopewash explain the presence of non-local cobbles and boulders. Especially important are the non-local angular, friable sandstone cobbles. These cobbles must have been transported into the area and deposited by a non-disruptive process. It is most likely that they were transported while frozen near the base of the ice and subsequently released passively by subglacial meltout.

The gradational contact with the overlying Unit 2 shows that both units are inter-related. Because Unit 2 (see full discussion next section) is glacigenic, Unit 1 must also be glacigenic. The preservation of the sandstone cobbles indicates that lodgement, a highly destructive process, could not have deposited the diamicton. Their preservation and the presence of intact sand and gravel lenses preclude the likelihood that Unit 1 is a deformation till that had been subjected to pervasive internal deformation. The unit is therefore either a glacially derived debris-flow deposit or a meltout till. Within the unit there is a vertical clast-gradation with locally derived shale phenoclasts dominant at the base and distally derived phenoclasts predominating near its top. The locally derived phenoclasts would have been eroded from the local shale when the ice was still active. The non-local phenoclasts were transported into the area. The preservation of this vertical gradation in phenoclasts would be best preserved by subglacial meltout of the debris-laden basal ice. The preservation of friable sandstone phenoclasts also favors deposition by passive subglacial meltout. The dense nature of the diamicton in places also suggests that the unit had a sub-glacial origin. The low eigenvalues of the pebble fabrics and the muddy gravels within the diamicton indicate that some alteration of the deposit has occurred. Ronnert (1989) interpreted a Swedish glacigenic diamicton as a subglacial meltout till, that during formation, had been locally altered by lodgement, internal deformation, and subglacial meltwater discharges. The sand lenses and dirty gravels beds, and gravel lenses composed of shale phenoclasts observed, indicates that Unit 1 may have been subject to similar modifications.

Unit 1 is therefore interpreted as a subglacial meltout till that has been locally modified by internal remobilization and meltwater discharges. It could be debated whether it should be called a meltout till but the dominant process for its formation was passive meltout. The subsequent alteration has reduced but not destroyed all the traces of its initial origin. The disruption and alteration of Unit 1 support the conclusion that subglacial meltout tills are commonly modified during their formation by other processes associated with immobile basal ice masses (Lawson 1979a, Paul and Eyles 1990, Shaw 1982, 1987).

Unit 2 Diamicton beds intercalated with clay and silt interlayers

Unit 2 conformably overlies Unit 1, is 1.5 to 5.0 m thick, and consists of thin very dark grayish brown (2.5Y3/2 dry) planar diamicton beds separated by continuous silt and clay beds (Fig. 3.13). There is a gradational contact between Units 1 and 2. Like Unit 1, the unit is thinnest (under 2m) at the knob's crest and thickens over the flanks (about 5m thick on the east and about 3 m thick over the west).

Diamicton beds

Individual diamicton beds, ranging in thickness from 3 to 10 cm thick and extending laterally for 3 to more than 10 m, have grain size distributions that are similar to the clayey diamicton of Unit 1. They also contain thin discontinuous stringers of silt (10-30 cm long) some of which contain angular silt intraclasts. Clasts comprise 5 to 10% of the volume of the unit but only 4 to 5% of all phenoclasts are erratic.

Silt and clay interlayers

The 1 mm to 3 cm thick interlayers can commonly be traced laterally for at least 10 m and consist of graded silt and massive clay laminae and beds, both of which contain numerous subangular silt intraclasts. Individual silt lamina, 0.1 to 0.5 cm thick, are usually continuous for more than 10 m. The number of silt interbeds decreases vertically. In some places the silt laminae and beds are composed of angular silt intraclasts within a silt matrix. In other areas, clay makes up 30% of the matrix of the laminae. Thin interlayers are loaded by overlying diamicton beds. Individual silt laminae are more irregular in thickness than are the clay interlayers. Thin, discontinuous diamicton stringers can be found within thicker silt interlayers.

Clay layers range from 0.5 to 2.0 cm in thickness. Some are traceable for about 10 m but most pinch-out rapidly over 2 to 3 m. Both the silt and clay interlayers conformably drape underlying diamicton beds and are commonly deformed beneath large phenoclasts. Some silt and clay interlayers drape large phenoclasts (Fig. 3.14). Fabrics measured in Unit 2 (Table 3.1) have low S_1 eigenvalues and have no preferred orientation.

Bed attitude in Unit 2 reflects the form of the underlying bedrock surface. On the east flank of the bedrock knoll the beds dip eastwards while on the west flank they dip westwards.

The genesis of the sediment in Unit 2.

Unit 2 is a complex deposit in which thin planar diamicton beds are intercalated with sand and silt laminae. Frakes and Crowell (1967) in their study of diamicton formation concluded that once the origin of the interbeds is determined, the genesis of the diamicton forming the bulk of the deposit is readily established. The silt interlayers are interpreted as underflow deposits on the evidence of their continuity, their normal grading and the sharp contacts between individual laminae. These are characteristic features of underflow deposits (Ashley 1975; 1989; Gustavson 1975). The silt interlayers may have been generated by periodic slumping of glacially-derived debris into the lake or by direct meltwater input. Silt laminae conformably overlain by thin clay layers are rhythmite couplets that can be deposited by distal deposition from slump-generated underflow currents or directly from meltwater inflows (Smith and Ashley 1985). The clay component of the couplets was deposited by suspension from the tail of turbid underflows or from the water column. These silt and clay couplets may be varves but at this stage there is no evidence to suggest annual deposition. It is better to classify the couplets as rhythmites since rhythmites carry no time connotation. Cases of several rhythmites being deposited in a single year are common (Gilbert and Shaw 1981; Smith and Ashley, 1985). Depositional of several rhythmites in a single year has been reported by Lambert and Hsu (1979) and Sturm and Matter (1978). The silt and diamicton intraclasts observed in the laminae could be ice-rafted or rain-out material. The angular silt and silt-clay intraclasts in some of the interlayers are either rip-up clasts eroded by strong underflows or remnants of rhythmites broken during slumping of stratified silt and silt-clay beds. The large phenoclasts and clusters of phenoclasts that deform underlying laminae and diamicton layers in Unit 2 are dropstones. Ovenshine (1970) believed that stone and stone clusters in association with silt and clay couplets and rhythmites characterized ice-rafted deposits. The silt laminae and the suite of structures associated with them could only have been deposited in a glaciolacustrine environment. The diamicton layers must therefore have also been emplaced subaquatically.

Two general depositional mechanisms have been advanced for subaquatic diamicton formation: (1) debris rain-out beneath a floating ice margin, (Gibbard 1980) or (2) debris-flow deposition near the front of ice ending in a lake (Evenson et al. 1977). The diamicton beds in Unit 2 are believed to be thin fluid debris-flow deposits. Fluid debris-flows, like those in Unit 2, are commonly thin (less than 10 cm thick), with nearly planar upper and lower bounding surfaces, and few phenoclasts (Bull 1964; Lawson 1979a). This would account for the scarcity of phenoclasts in the diamicton layers and the high proportion of matrix (silt and clay grain sizes) observed in thin-sections of these diamicton layers. In contrast, viscous debris-flows have irregular upper and lower surfaces, a wide variety of geometries, and an abundance of phenoclasts (Lawson 1979a; Ghibaudo 1992). An iceberg origin for the planar layers is unlikely as deposits from icebergs are generally identified as discontinuous, irregularly-shaped mounds of poorly sorted material (Thomas and Connel 1985). The thin, continuous, planar nature of the diamicton beds does not fit this description. A rain-out origin for the diamicton layers is difficult to explain. The planar continuous nature of the diamicton layers would require a large uniform concentration of debris at the base of the ice. This uniformly-distributed debris would then have had to settle through the water column without being affected by currents. Measurements of debris concentration and distribution at the base of modern glaciers and ice caps (Boulton 1967, 1968, 1975; Hooke 1970; Holdsworth 1973a,b; Lawson 1979a) indicate that the debris is neither highly concentrated nor is it uniformly distributed. Instead debris layers are highly irregular and commonly contain only a few per cent debris per unit of debris-laden ice. The abundance of rhythmites and silt laminae intercalated with the diamicton layers of the unit indicates that turbid underflow currents were active in the lake. This would tend to prevent even dispersal of debris as it settled through the water column thereby preventing formation of thin planar diamicton beds as observed in the unit. A rainout origin for the diamicton layers is therefore unlikely. Furthermore, silt intraclasts within the diamicton beds and disruption of the silt layers beneath diamictons beds are most likely the result of overriding and erosion of silt laminae by debris-flows. Rain-out and iceberg sedimentation would not erode or disrupt the silt laminae in a similar fashion.

The deposition of the diamicton beds was complex and may have been a two stage process. Initially, debris, released by rainout, accumulated beneath an ice overhang. This sediment was mixed with proximal debris-flows and suspension deposits. This water-saturated debris was unstable. It failed and thin, fluid debris-flows flowed into the deeper parts of the lake. The debris-flows may have mixed with lake water to form thin density currents that in turn, deposited some of the silt and clay interlayers. Inflows of turbid water and water column suspension sedimentation also contributed to the formation of the final deposit.

The conformable attitude of the diamicton beds and interlayers of Unit 2 and the underlying bedrock surface can also be explained. If Unit 1 is a subglacial meltout till, as has been suggested, it is possible that the diamictons and laminae of Unit 2 were originally deposited horizontally over stagnant, debris-laden ice. Melting of the buried debris-laden ice produced the subglacial meltout till of Unit 1 and brought Unit 2 (composed of diamicton beds and silt and clay laminae) into contact with Unit 1. The melting of a greater thickness of ice on the flanks of the bedrock knob than at its crest, caused the originally sub-horizontal strata of Unit 2 to become inclined with a final resting attitude reflecting the underlying bedrock surface.

Extent of Lake Cypress

The size of Lake Cypress is difficult to estimate as its southern margin was formed by ice and there are no well-defined beaches or deltas. A conservative reconstruction (Fig. 3.1B) using the Cypress Lake sections as the approximate location of the south shore of the lake, the bedrock highs where the Frenchman channel narrows to the west and east as the east margin, and the north wall of the Frenchman channel as the northern margin, indicates that a lake of about 56 km^2. If the water surface was at the top of the north wall of the channel (1090 m), the lake would have been about 90 m deep. The lake therefore would have contained between 4 x10^9 and 5 x10^9 m^3 of water. This may underestimate the volume of the lake as it is possible that the lake occupied a much larger area between the East and West lobes and may have extended a considerable distance onto the ice to the south, making it a partially supraglacial lake. This is especially true if the intermixed moraine and lake sediment (unit Mix of Klassen 1991, 1992) which occurs as a band around most of the southern flanks of the East and Center blocks wee deposited in this lake.

Discussion

A. Local implications of lake impoundment

The impounding of Lake Belanger in a sidehill position is evidence that ice encroached onto the south slopes of the East and Center blocks. Since the sediment deposited in Lake Belanger covers meltwater channels that extended south from the Late Wisconsinan ice mass north of the Cypress Hills, the lake must also have been Late Wisconsinan. Visual inspection of deposits derived from the East Lobe and thin sections from these deposits indicates that the commonly lack shale pebbles whereas deposits derived from the West Lobe are characterized by abundant rounded and sub-rounded shale pebbles. The absence of shale pebbles in the glacial deposits in the Canal and Gilchrist sections suggests that it was the East Lobe that impounded the lake.

To have ice of the East Lobe on the south flanks of the East Block requires a more extensive East Lobe than is depicted in Christiansen (1979), Clayton and Moran (1982), Dyke and Prest (1987a), and Klassen (1989). A more extensive East Lobe (Klassen 1992 and this paper) requires that the distribution of all the ice lobes around the Cypress Hills at the Late Wisconsinan maximum, be reconsidered. Placing the East Lobe onto the southern slopes of the East Block also means that the Frenchman channel cannot be an interlobate meltwater channel as previously depicted (Christiansen 1979; Clayton and Moran 1982; Dyke and Prest 1987a) because the East Lobe covers the area of the channel. The alternative origin advanced in Chapter 4 of this volume, is that the channel was incised during the early stages of deglaciation, by catastrophic release of meltwater impounded on and around the southern flanks of the Cypress Hills. Glacial lakes Belanger and Cypress lay east of the head of the Frenchman channel at Merryflat. Catastrophic release of these two lakes therefore could not have cut the initial reach. An additional lake at the Merryflat area is therefore needed (Chapter 4, this volume). Lakes Belanger and Cypress however, did drain through the Frenchman channel and played a role in the incision of the Frenchman channel to the east.

Lake Belanger infilled the low between the East and Center blocks to an elevation of about 1125 m above sea level (Fig. 3.1B). The lake occupied much of the terrain that was designated as the "residual drift landscape complex" and "bedrock with residual drift landscape complex" (Klassen 1992), (Fig. 3.15). An alternative, simpler hypothesis is that the isolated surface erratics that lie northward and upslope of the limits of deposits and landforms directly attributable to the Late Wisconsinan ice, were deposited by iceberg rafting across Lake Belanger. The erratics limit, therefore, coincides with the maximum level of the lakes impounded on the southern flanks of the Cypress Hills. An older more extensive ice sheet is not needed to explain the distribution of these erratics.

B. A glaciolacustrine depositional sequence for the Prairies.

In lakes Belanger and Cypress, an unbedded diamicton unit graded into an overlying 'stratified glaciolacustrine diamicton' unit. A similar sequence was observed in sediment deposited in an unnamed glacial lake in central Alberta (Kulig 1985) and in gravel pits in the Edmonton area (Kulig 1989; Shaw 1982). In all of these sections, the underlying sedimentary units had gradational contacts with the overlying units. Additionally diamicton beds within each sedimentary unit were identical in grain size, color, compaction, and types of erratics and therefore could not be differentiated by these parameters. Only the interlayers, intercalated with the diamicton beds, enabled the exposures to be sub-divided and the genesis of the deposits to be determined. In Lake Belanger and central Alberta (Kulig 1985), the unbedded diamicton units were interpreted as consisting of coalesced debris-flows. The absence of intercalated sand and silt laminae probably reflects a short time interval between debris-flow events and/or a lack of basal currents. Thus there would have been insufficient time between debris-flow events or insufficient sedimentation from turbid underflows to form a recognizable interlayer. At Cypress Lake, the underlying unit was interpreted to be a subglacial meltout till that had been strongly modified in places by meltwater activity.

Formation of similar glacial lake sequences in other areas requires several factors: (1) debris-laden ice terminating in a glacial lake, (2) deposition of this debris at the grounding line, (3) ice preserved at or near the base of the lake. The basic components of the depositional model have been observed in several modern settings. Holdsworth (1973a, b) and Barnett and Holdsworth (1974) described the eastern margin of the Barnes Ice Cap which is grounded in Generator Lake . The ice cap terminates in 30 to 60 m of water and has large undercut ice overhangs. Deposition of debris occurs at or near this grounding position (Holdsworth 1973a, b; and Barnett and Holdsworth 1974). Gustavson (1975) reported extensive preservation of ice beneath lake sediment in Malaspina Lake. The depositional sequence would be as follows. Melting of ice in an ice-contact lake occurs rapidly at the ice-water interface by thermal and wave erosion. A notch is formed in the ice wall, with an underlying ice apron and an upper ice overhang

(Fig. 3.16A, B). Debris is released into the lake by rainout from the ice overhang and by debris-flow activity originating from the area of the notch and the main ice mass. This sediment accumulates over the underlying ice apron, insulating it from rapid melting. Periodically pieces of the ice overhang break off and this debris-laden ice is embedded in the sediment accumulating over the ice apron. The extent of the ice overhang would be reduced by brash ice calving (Sugden and John 1976). This releases numerous small icebergs that can raft fine sediment and boulders throughout the lake. Brash ice calving also releases a large volume of sediment at the calving line (Sugden and John 1976, p. 226). This unstable water-saturated debris builds up over the underlying debris-laden ice apron partially insulating it. When this unstable sediment fails it is transported further into the lake basin by debris-flows (Gustavson 1975), some of which are very fluid in nature, or by slumps that subsequently change to sediment flows (Morgenstern 1967; Hampton 1972). These slumps can also generate underflow currents (Hampton 1972). These underflow currents as well as those generated by turbid inflowing streams can deposit sand or silt laminae and beds over the debris-flow surfaces. They can also winnow the top of the debris leaving a sand lag (seen in the sections along Battle Creek in central Alberta). Weaker underflows in more distal locations would deposit thin, normally-graded silt laminae with sharp contacts between them, (observed in the central Alberta and the Cypress Hills sections). Rapid retreat of the ice overhang could result in an extensive buried ice apron along the bottom of the lake. When this underlying debris-laden ice apron melts, a meltout till, with irregular boundaries, is formed. In places this meltout till can be greatly altered both during and after deposition, to form a modified meltout till like the one observed at the base of the Cypress Lake section. Waterlain debris flows and suspension and underflow sediments are lowered onto the meltout deposit as the ice apron melts. Meltout of the underlying debris-laden ice can cause the overlying interbedded debris flow and underflow sediments to become reoriented as observed for Unit 2 in the Cypress Lake section. Superimposition of waterlain diamicton over subglacial tills has been reported by Shaw (1982, 1987) and Dreimanis (1989). The typical sequence would be a subglacial meltout till of variable thickness and lateral extent overlain by a glaciolacustrine debris-flow complex. If there were subglacial and englacial meltwater conduits, the meltout till could, in places, be so completely changed by meltwater action that it is no longer a meltout till. Remobilization of the water saturated subglacial sediment could also occur generating subaquatic glacigenic debris-flows.

A second possible environment for superimposing glaciolacustrine diamicton deposits over meltout tills would be in a supraglacial lake setting. The lake forms on top of the ice and not in front of it. The sedimentary sequence produced would be very similar to that for ice fronting in a lake. This is especially true if the buried ice apron in the ice contact lake was extensive. One characteristic of a supraglacial lake would be the absence of beaches and well-defined shore-lines. It is likely that many of the glacial lakes that covered Alberta and Saskatchewan were at least partially supraglacial.

Since the retreat of the last ice invariably impounded lakes along its margin, the sequence of glaciolacustrine sediments overlying meltout tills should be commonplace on the Prairies. Shaw (1982, 1987) and Feltham (1987) described similar situations where meltout till is overlain by glaciolacustrine diamicton. Unit 2 in the Canal section and the unbedded diamicton of Kulig (1985), interpreted to be debris-flow units from their fabrics and their association with overlying glaciolacustrine diamicton units, could be meltout till that was sufficiently remobilized for its phenoclasts to be re-oriented. The meltout till-glaciolacustrine diamicton sequence observed at the Cypress Lake sections would therefore be present. In areas where the basal-ice apron is thin, absent, or lacks debris, the subglacial meltout till would be thin or missing. The absence of the subglacial meltout till could also reflect deposition of the meltout till in an unstable location (such as on a slope). Total or partial remobilization of the meltout sediment would result in a sedimentary sequence that began with a lower debris-flow unit.

The occurrence of this sequence in several areas in Alberta and Saskatchewan supports one conclusion in Chapter 2 of this volume, that the impounding of ice-contact lakes invariably leads to the deposition of interlayered diamicton layers and sorted-sediment beds and laminae. In areas where drainage is not impounded, the expected sequence is subglacial till, overlain by a variable thickness of subaerial glacigenic debris-flows and glaciofluvial sediment. Since the proportion of diamicton in glaciolacustrine diamicton complexes is large and the intercalated laminae and beds are easily mistaken for shear planes or preserved englacial features, glaciolacustrine diamicton units are easily misidentified, especially in borehole investigations, as subglacial tills or subaerial debris-flows. The result is that the deposit is mistakenly identified as a subaerial till. The distribution of glaciolacustrine sediments and the extent of glacial lakes would therefore be greatly underestimated. Odynsky and Newton (1950) and Odynsky et al. (1952) described lacustro-tills over much of central Alberta and glaciolacustrine diamictons have been described by May (1977), Proudfoot (1985), and Liverman (1989) but the volume of mis-identified glaciolacustrine sediment is unknown. If the extent and volume of this sediment have been greatly under-estimated, the number and more importantly the extent of these glacial lakes on the Prairies during the last glaciation would also have been underestimated. this underestimation in turn would lead to underestimation of the role and affects that the draining of these extensive glacial lakes had on the formation of the Prairie landscape.

Conclusions

1) Stratified diamicton units, on or near the south slopes of the East and Center blocks of the Cypress Hills of Saskatchewan, were deposited in ice-contact lakes. These deposits are very similar to other exposures on the Prairies interpreted as glaciolacustrine diamicton deposits (Kulig 1985; Proudfoot 1985) and support the viewpoint (Chapter 2, this volume) that impounding of ice-contact lakes **always** leads to the deposition of such sediments and that the volume of stratified glaciolacustrine diamicton deposits have been greatly underestimated.

2) The impounding of Lake Belanger required the East Lobe to cover the area of the Frenchman channel and ascend onto the southern flanks of the East Block (Klassen 1992). The Frenchman channel therefore, could not be an interlobate, meltwater channel formed by gradual fluvial erosion. Catastrophic incision of the channel by sudden drainage of glacial lakes is hypothesized in Chapter 4 of this volume. Since lakes Belanger and Cypress lay east of the head of the channel at Merryflat, a large glacial lake must have been impounded in that area.

3) Ice rafting of erratics northward across Lake Belanger explains the distribution of erratics to 1125 m elevation and beyond the limits of glacial deposits and glacial landforms from the last ice. Meltwater channels draining southward from the northern ice mass also carried erratics across the Cypress Hills. These erratics are located in or nearby the meltwater channels that flowed from the northern ice mass. These two mechanisms explain the distribution of erratics beyond the limits of the Late Wisconsinan ice and account for the erratics described by Klassen (1992) in his "bedrock with residual drift landscape complex".

Table 3.1. Phenoclast fabric data for fabrics measured in the Canal, Gilchrist and Cypress Lake sections.

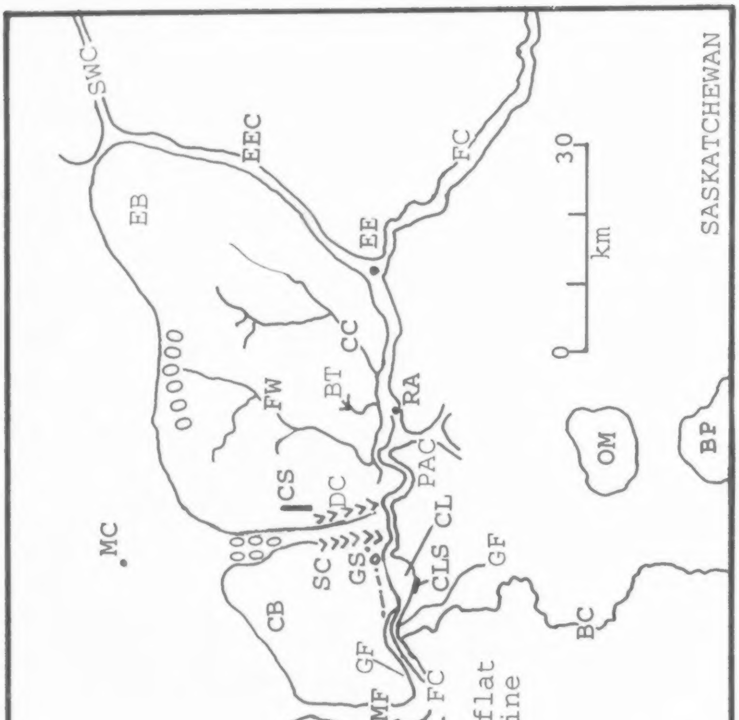
FABRIC NO.	UNIT	TREND	PLUNGE	S_1	S_2	MULTI-MODAL	CLAST NO.
CS1-89	2	76.3	9.2	0.48	0.41	Y	26
CS2-89	2	63.6	6.5	0.46	0.35	Y	26
CS3-89	2	53.7	18.2	0.51	0.38	Y	25
CS1-90	3	317.3	6.4	0.64	0.27	Y	30
CS2-90	3	298.9	1.0	0.67	0.23	Y	31
GIL1-89	3	211.2	2.7	0.83	0.14	N	24
CL1W	2	339.4	12.4	0.59	0.39	Y	22
CL2W	2	340.4	14.1	0.43	0.34	Y	30
CL3W	2	346	15.9	0.68	0.26	Y	30
CL6	3	337.6	11.3	0.60	0.37	Y	30
CL2E	3	165.8	9.4	0.48	0.43	Y	30
CL4E	3	160.9	10.0	0.52	0.38	Y	30
CL6W	3	177.1	13.3	0.48	0.30	N	31

Figure 3.1A. Location map showing places mentioned in the text. BAC- Battle Creek, BP- Boundary Plateau, BT- Blacktail Creek, CB- Center Block, CC- Conglomerate Creek, CL- Cypress Lake, CLS- Cypress Lake Sections, CS- Canal Section, DC- Davis Creek Coulee, EB - East Block, EE- Eastend, EEC- Eastend Coulee, EL- Elkwater, FC - Frenchman Channel, FW- Fairwell Creek, GS- Gilchrist Section, MC - Maple Creek, MCC- Middle Creek Channel, MF- Merryflat, OM- Old Man On His Back Plateau, PAC- Palisades Coulee, RA- Ravenscrag, SWC- Swift Current Creek, SC- Sucker Creel Coulee, UC- Underdahl Channel, WB- West Block. GF marks the location of a glaciofluvial delta along the south flank of the Center Block and outwash at the west end of Cypress Lake.

xxxxx	end moraine
⬭	Merryflat end moraine
•·•·•·	erratics limit
00000	ice-pushed gravel
—·—·—	discontinuous meltwater channels
>>>>>	coulees

3.1B. The location of lakes Belanger and Cypress and the distribution of the three ice lobes around the Cypress Hills during the early stages of Late Wisconsinan deglaciation.

//////	coalescence zone of the East and West ice lobes
·•·•·•·	erratics limit
≈	ice
⬭	ice-covered upland
⊥⊥⊥⊥	ice margin
≈≈≈	glacial lake
- - -	maximum limit glacial Lake Cypress
➞	water flow direction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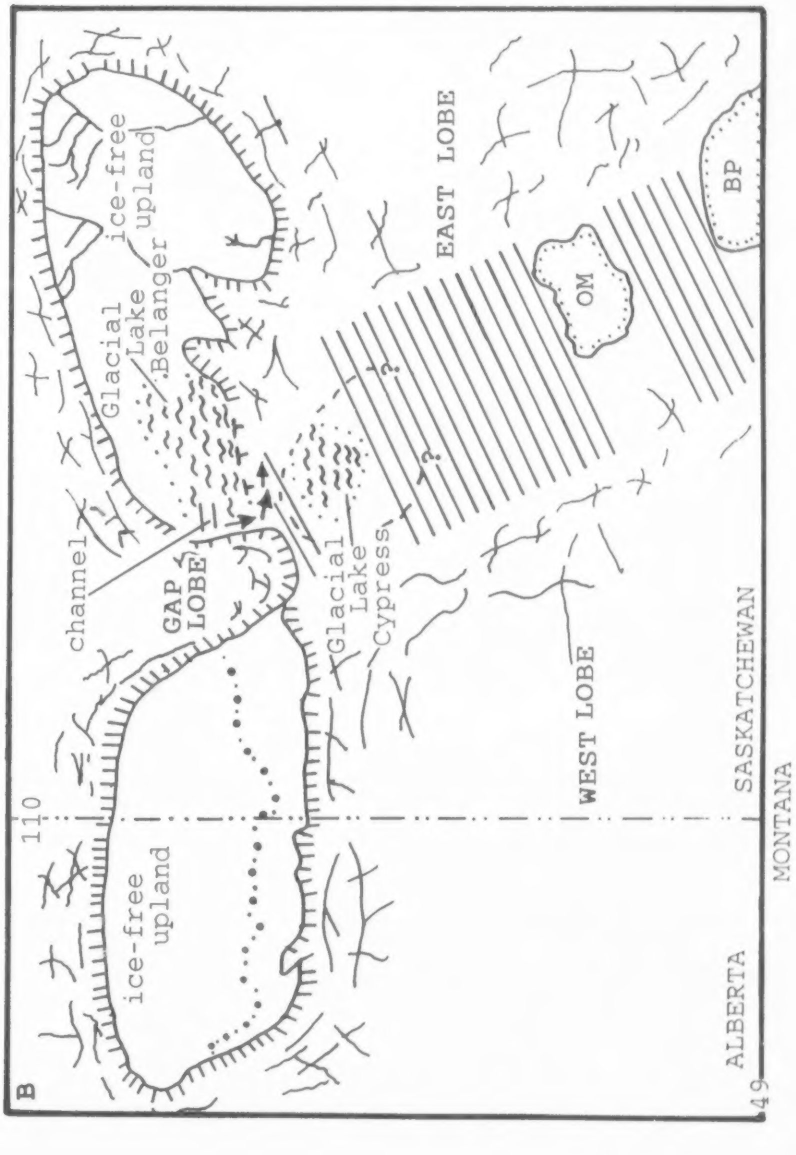
B
ice-free upland
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channel
GAP LOBE
Glacial Lake Belanger
ice-free upland
Glacial Lake Cypress
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EAST LOBE
WEST LOBE
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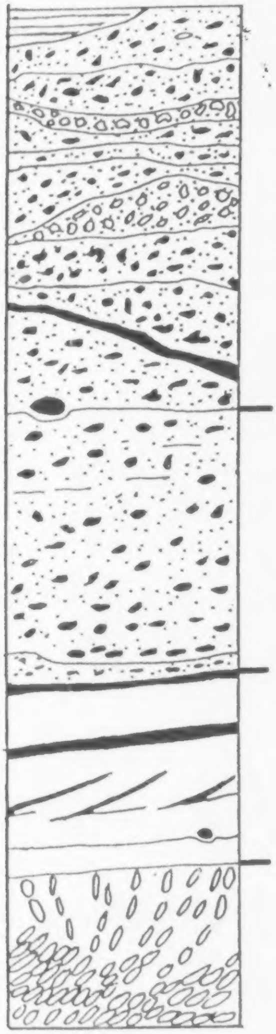
BP
SASKATCHEWAN
MONTANA
ALBERTA
49

Figure 3.2. Composite section of the Gilchrist and Canal Sections.



Unit 3 Diamicton interbedded with silt and clay layers. Dropstones and clay layers drape the underlying diamictons. The volume of clay interbeds increases upsection.

Unit 2. Massive appearing diamicton with gradational upper and lower contacts. Clasts at the base of the unit have sub-horizontal a-b axial planes.

Unit 1: Silt and clay rhythmites with thrusts and dropstones. The unit overlies the Cypress Hills Formation unconformably.

Cypress Hills Formation, preglacial boulders, gravels and sands, periglacial involutions and vertically oriented phenoclasts present in the upper 2 meters.

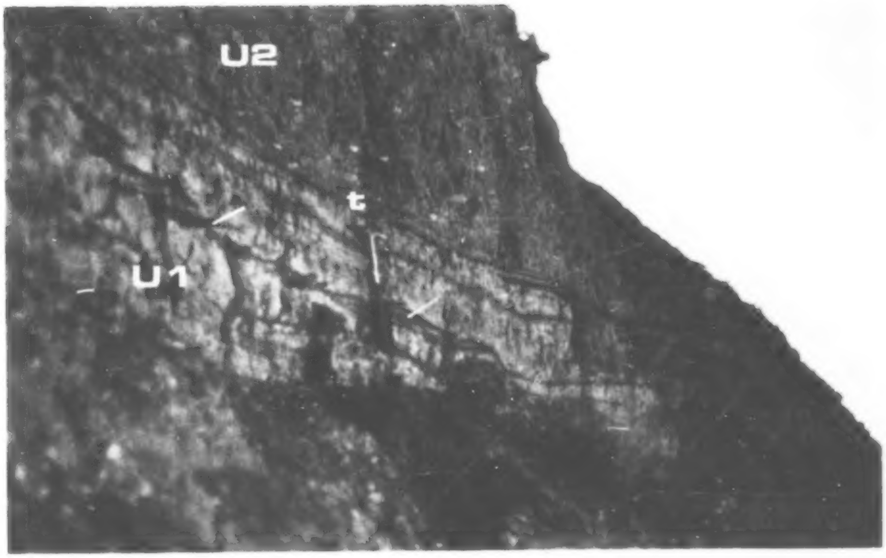


Figure 3.3. Photograph of the Gilchrist section showing the lower normally-graded silt unit overlain by the unbedded diamicton unit. The transition zone between the two units is marked t. The sheared clays in Unit 1 are indicated by white dashes. The pick is 75 cm long.

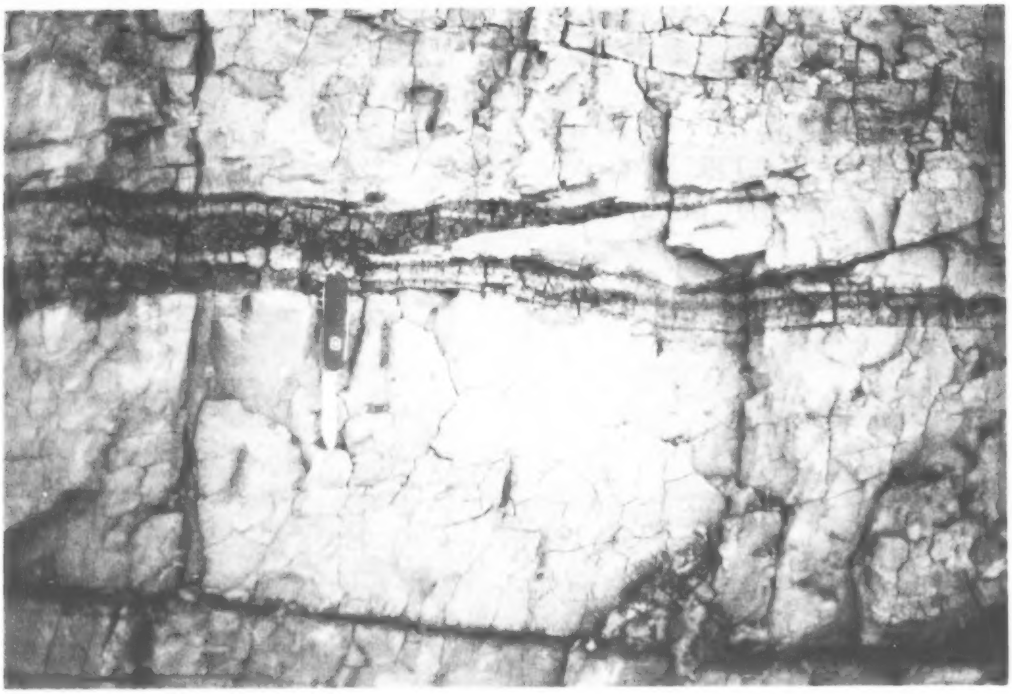


Figure 3.4. Close-up of the deformation of clay components of the rhythmites in the center of Unit 1 of the Gilchrist Ranch section. The deformations trend to the north which is upslope. Knife is 16 cm long.

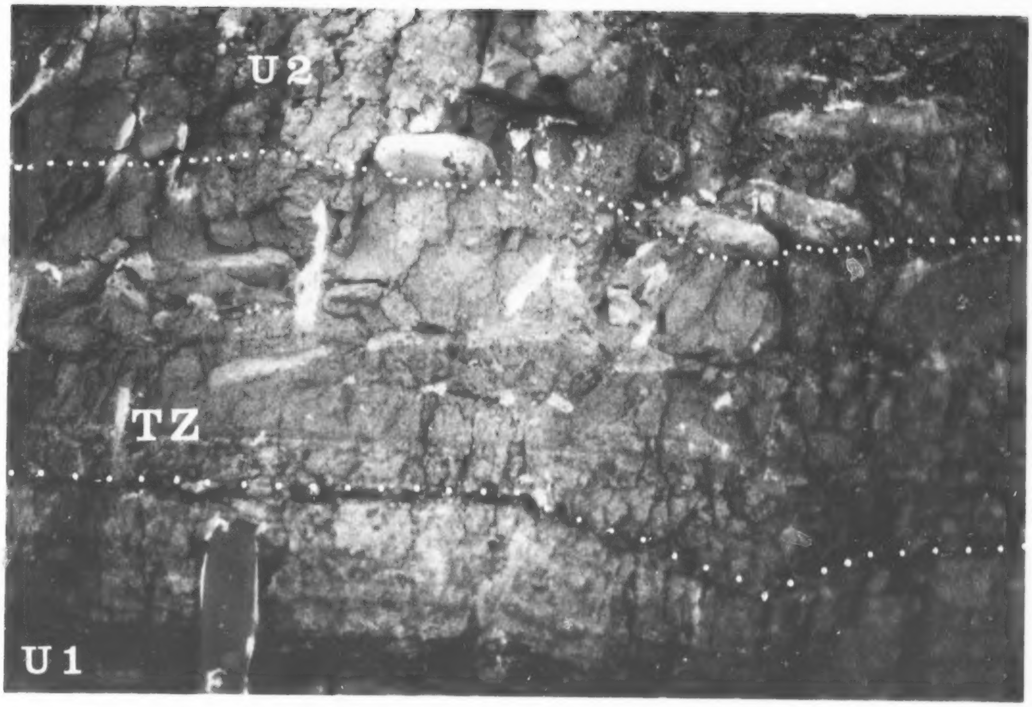


Figure 3.5. Close-up of the gradational sequence between Units 1 and 2 in the Gilchrist Ranch section. The contact between units 1 and 2 is indicated by the dotted line. The gradational sequence between them is labeled TZ. Note the horizontal a-b axial orientation of the phenoclasts in the photograph as discussed in the text. Knife handle is 8 cm long.

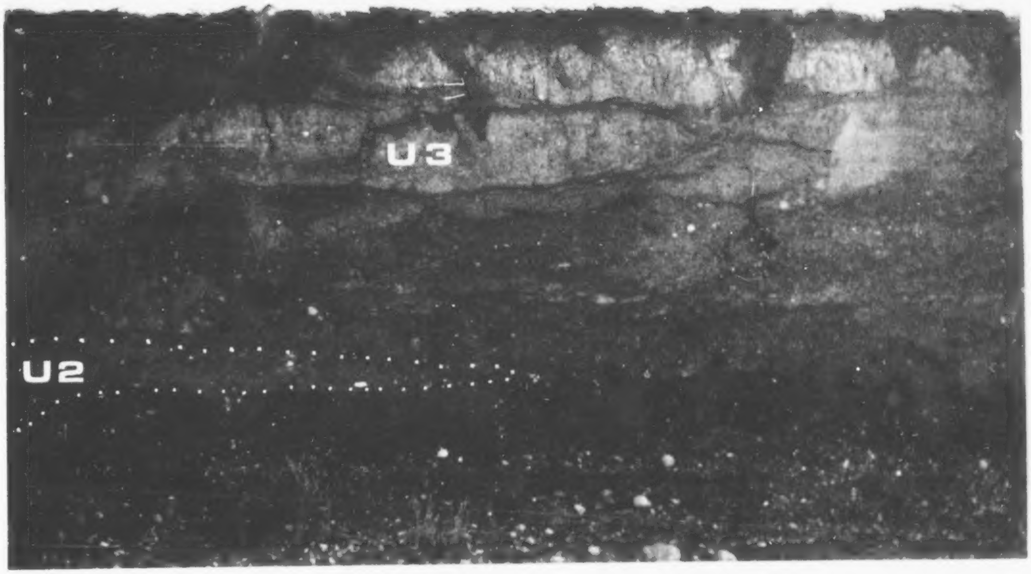


Figure 3.6. Photograph of units 2 and 3 in the center of the Canal section. The massive unit 2 is overlain by the stratified unit 3. The contact is marked by a white dotted line. The volume of clay beds in unit 3, increases upsection. In restricted locations, the clay beds can make up 90% of the unit's volume. Near the top of the section diamicton layers composed of angular silt intraclasts are interbedded with the clay beds 1 to 3 cm thick. Pick handle is 75 cm long.

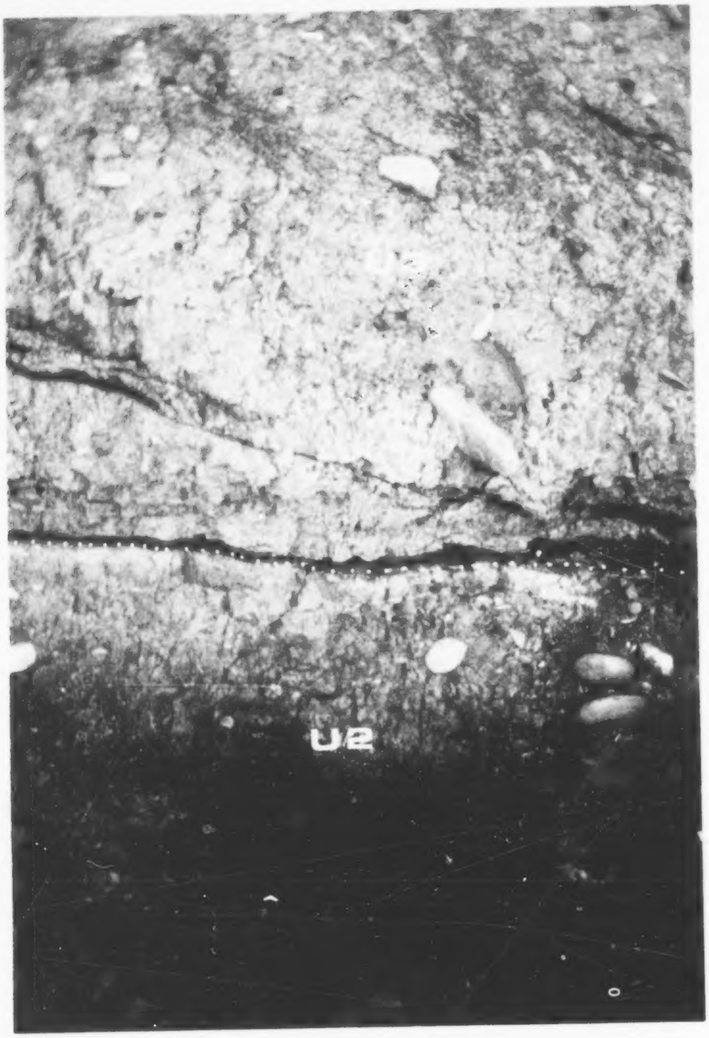


Figure 3.7. Close-up of Units 2 and 3 in the Canal section. The unbedded Unit 2 is overlain by a clay bed (just above the dotted line) which continues for over 40 m. Note the attitude of the upper clay bed and lens-like geometry of the diamicton layer immediately above the sub-horizontal clay bed. Knife handle is 8 cm.

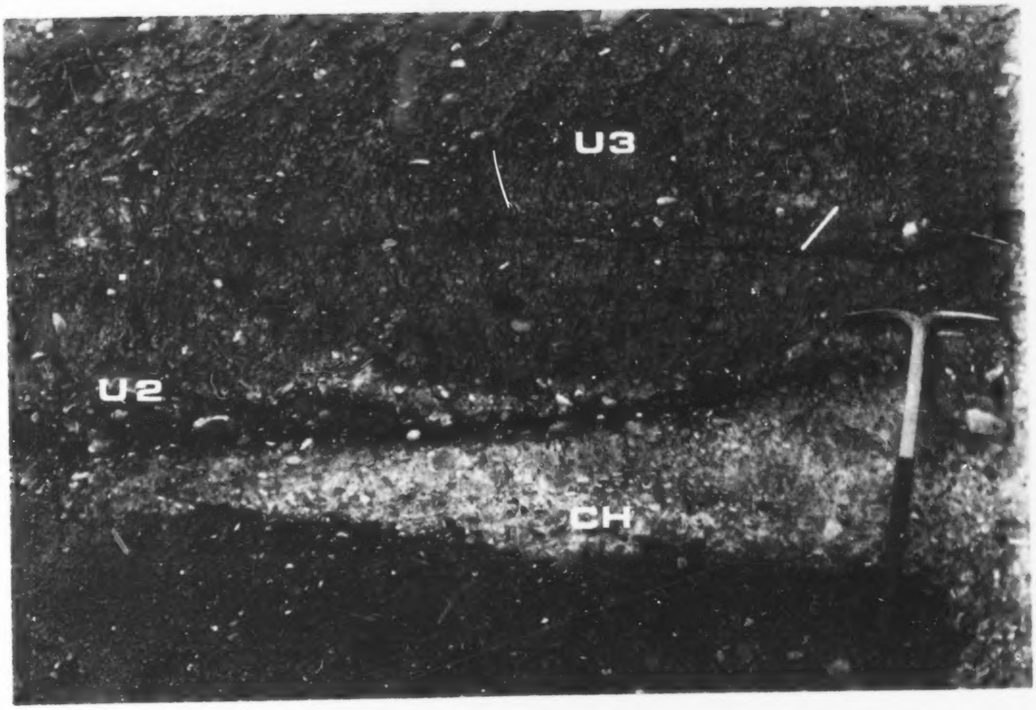


Figure 3.8. Photograph showing a continuous sub-horizontal clay bed (white dash) overlying a thin diamicton layer in Unit 3 of the Canal section. Unit 1 is absent and Unit 2 lies directly over the Cypress Hills Formation (CH). Pick is 75 cm long.

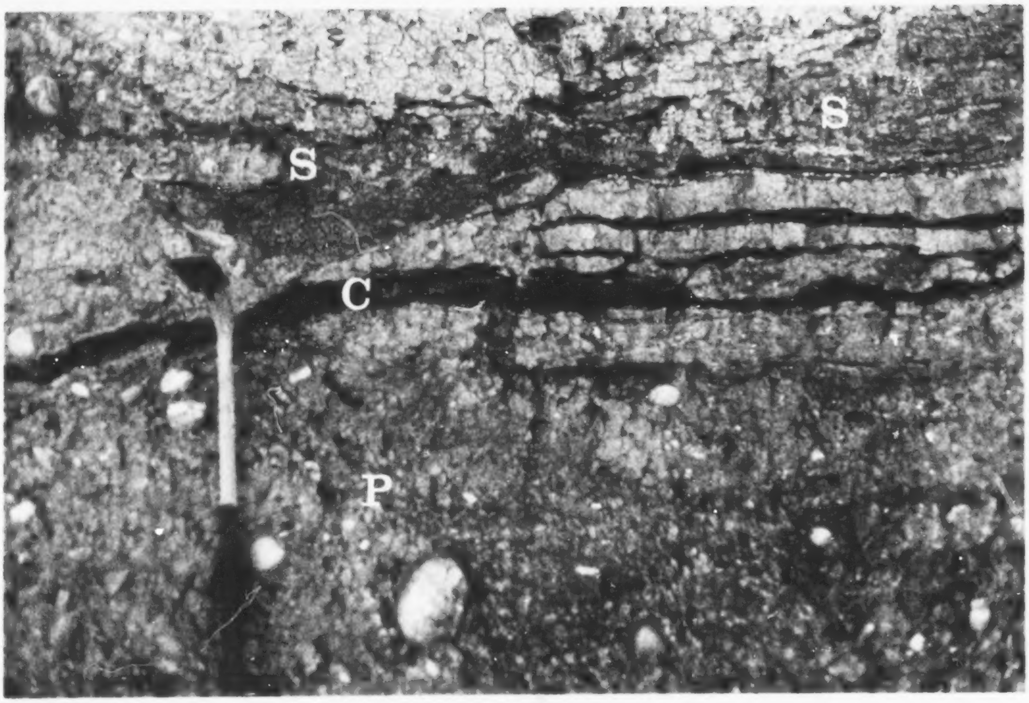


Figure 3.9. Photograph showing a phenoclast-rich diamicton layer (P) overlain by a clay bed (C) and a diamicton layer composed of normally graded silt intraclasts (S) in the Canal section. Pick handle is 75 cm long.

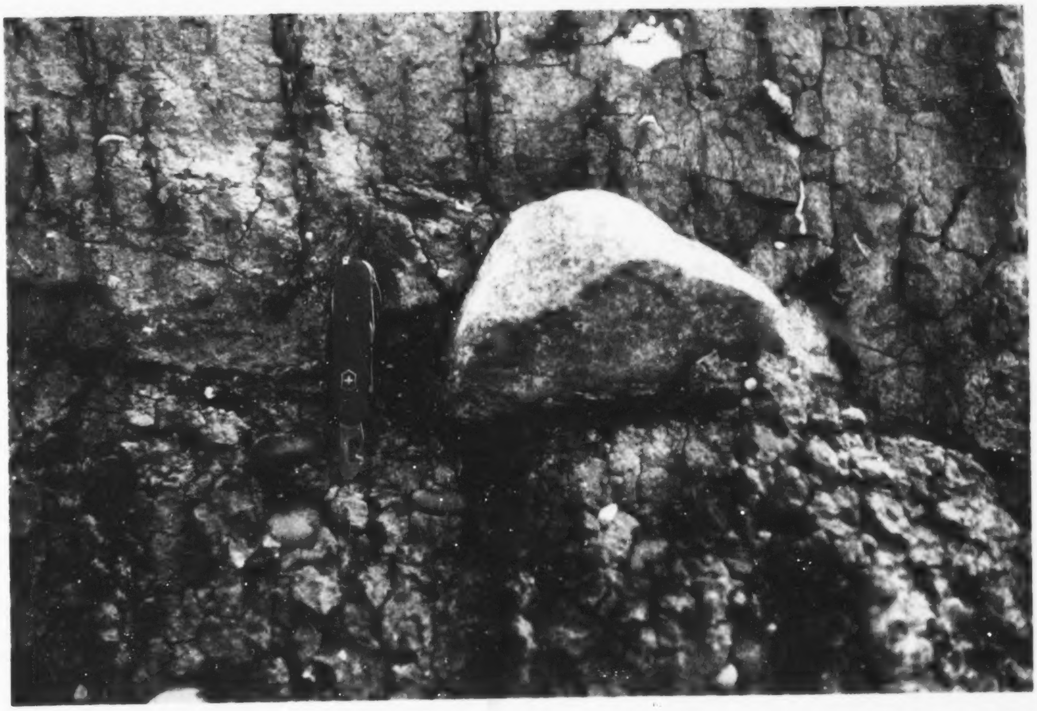


Figure 3.10. Dropstone deforming clay bed and underlying diamicton surface in Unit 3 of the Canal section. Knife handle is 10 cm long.

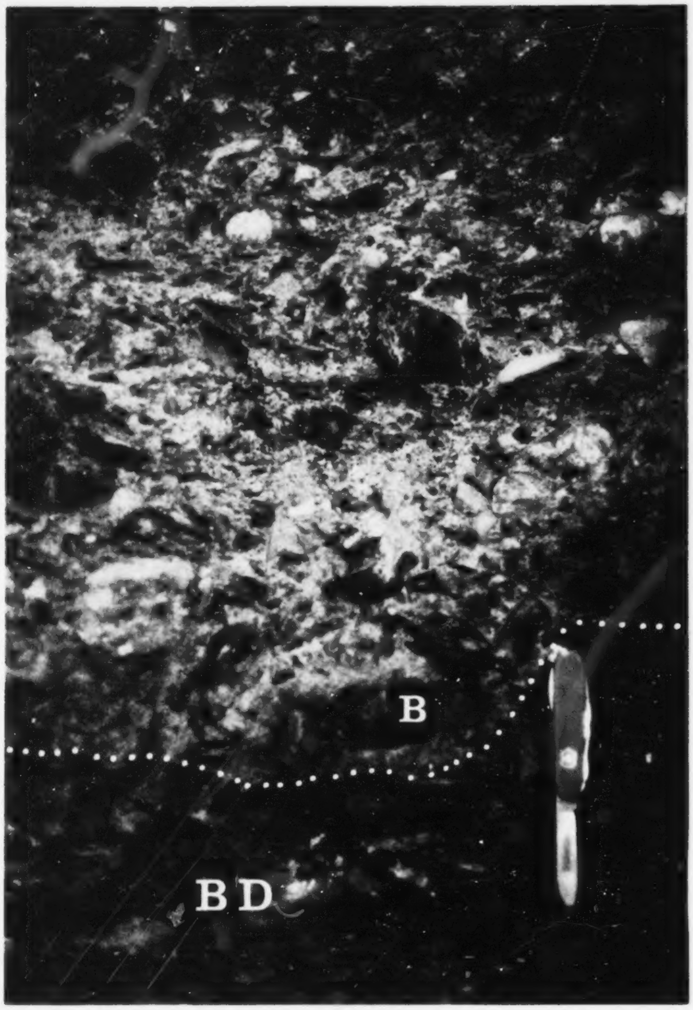


Figure 3.11. Contact between the shale (BD) and the overlying sandy-gravelly diamicton of Unit 1 of the Cypress Lake section. Note the irregular nature of the contact (white dots) and the abundance of angular shale phenoclasts such as B, derived from the underlying shale in the diamicton. Knife handle is 10 cm long.

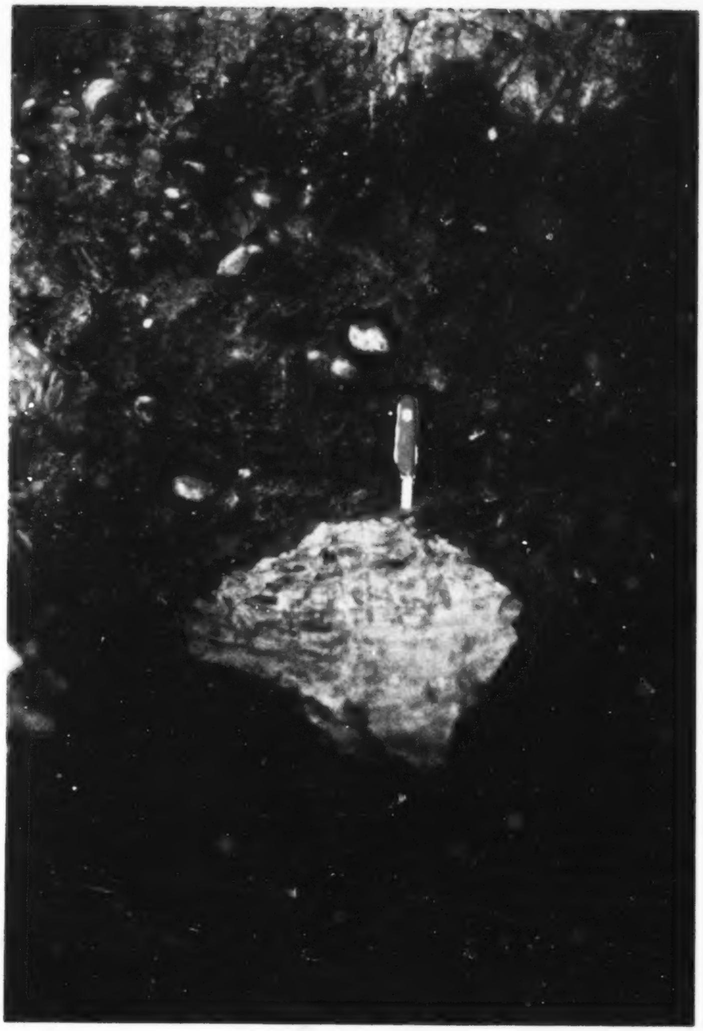


Figure 3.12. Photograph of a large angular friable sandstone boulder contained within the diamicton of Unit 1 of the Cypress Lake section. Knife handle is 10 cm long.

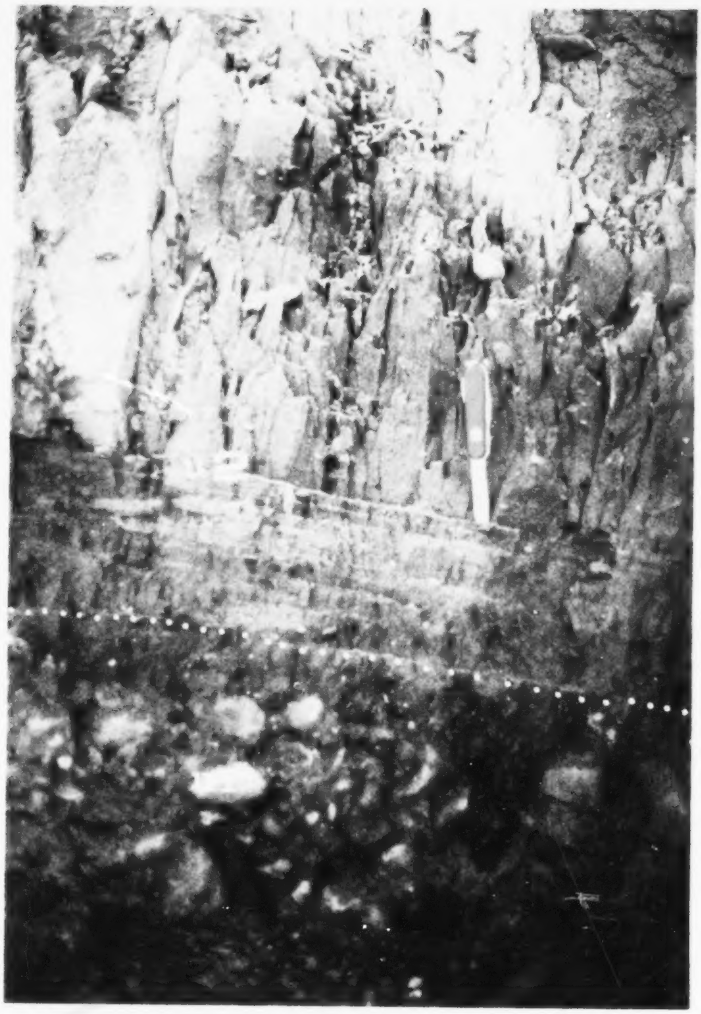


Figure 3.13. Contact between the diamicton of Unit 1 and the stratified diamicton of Unit 2 on the west flank of the bedrock high in the Cypress Lake section. The contact is indicated by white dots. The lower 15 cm of the diamicton in unit 2 is very finely stratified. Note the dip of the strata to the right. Strata on the east flank of the outcrop dip to the left. Here, the diamicton of unit 1 is very gravelly. Knife handle is 10 cm long.

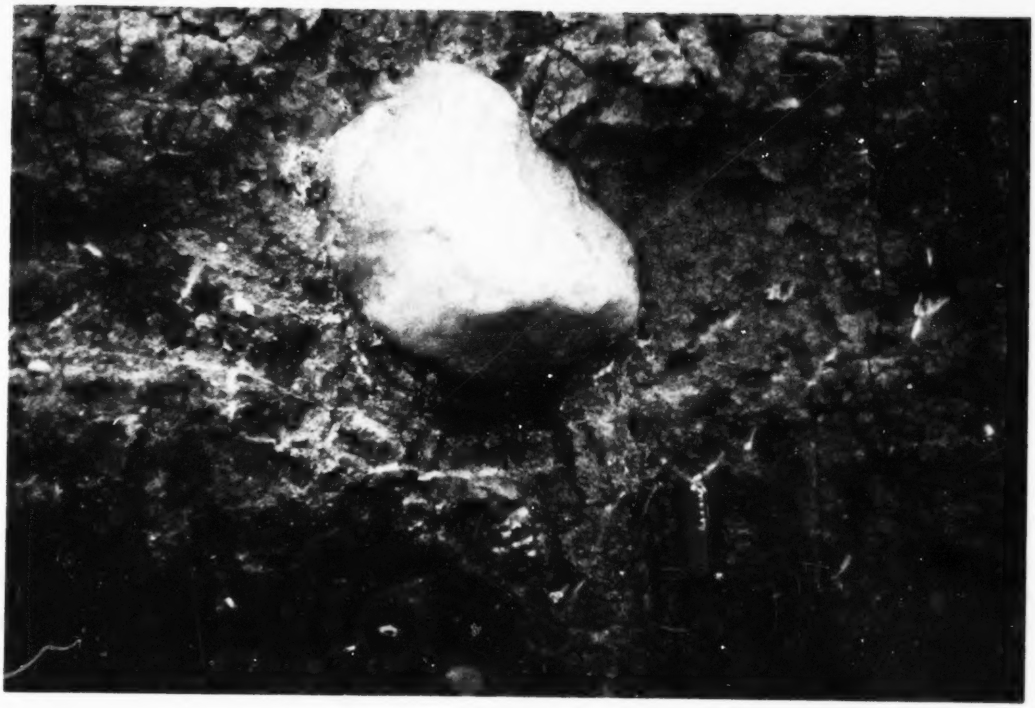


Figure 3.14. Deformation of silt laminae and diamicton layers in Unit 2 of the Cypress Lake section beneath a boulder (dropstone). Knife handle is 10 cm long.

Figure 3.15. Diagram showing the three landscape complexes: the "bedrock terrain with residual drift", "bedrock terrain with drift" and "first advance drift" described by Klassen (1992). Klassen (1992) identified the "bedrock terrain with residual drift" as Nebraskan or Kansan, the "bedrock terrain with drift" as Illinoian or Early Wisconsinan and the "first advance drift" as probably the deposit of an unspecified southwest ice lobe. The study area is enclosed in the box. Letter Code: BP- Boundary Plateau, FC- Frenchman Channel, LE- Lethbridge, MH- Medicine Hat, MKR- Milk River Channel, MR- Missouri River, OM- Old Man On His Back Plateau, SC- Swift Current, WM- Wood Mountain Upland. 1- West Block, 2- Center Block, 3- East Block.

░ bedrock terrain with residual drift landscape complex

▨ bedrock terrain with drift landscape complex

▥ first advance drift landscape complex

□ study area

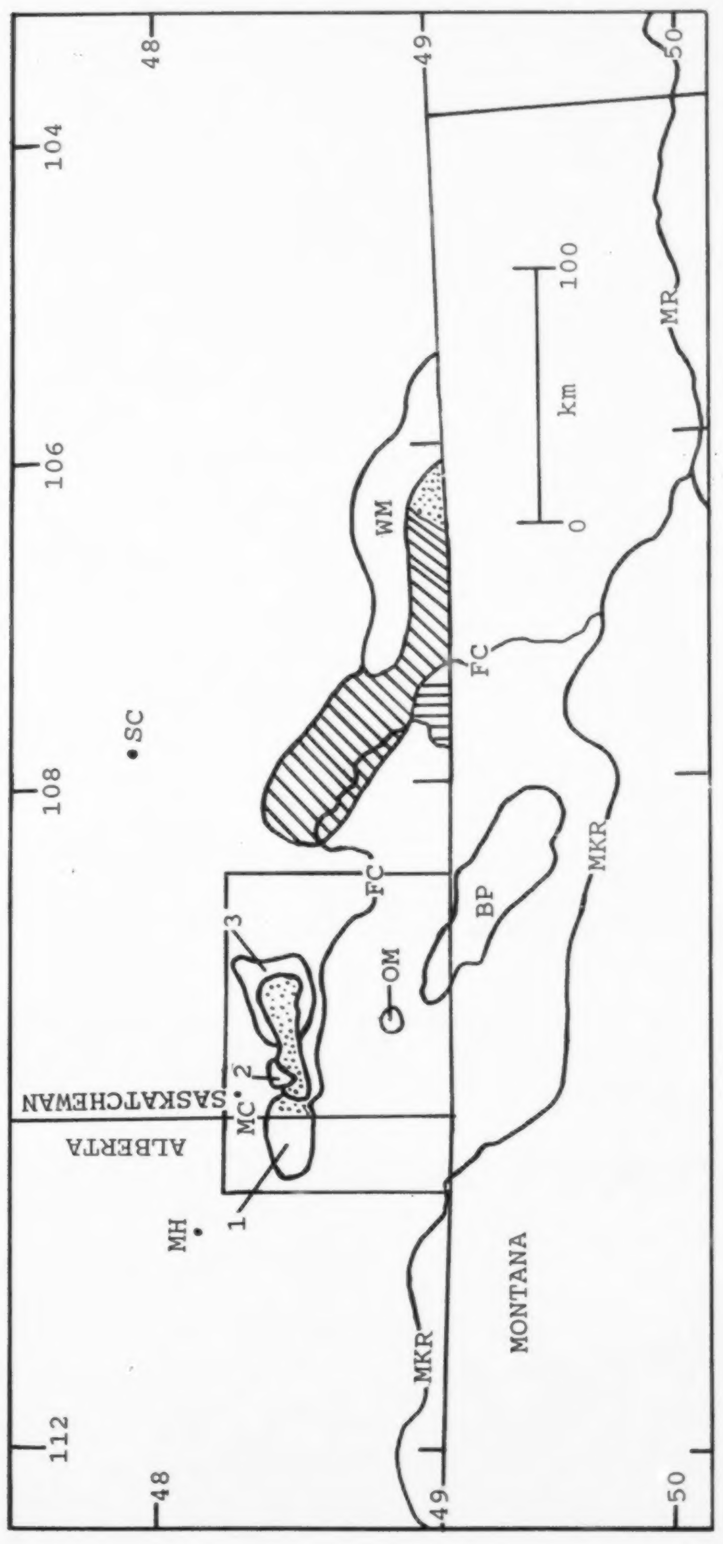
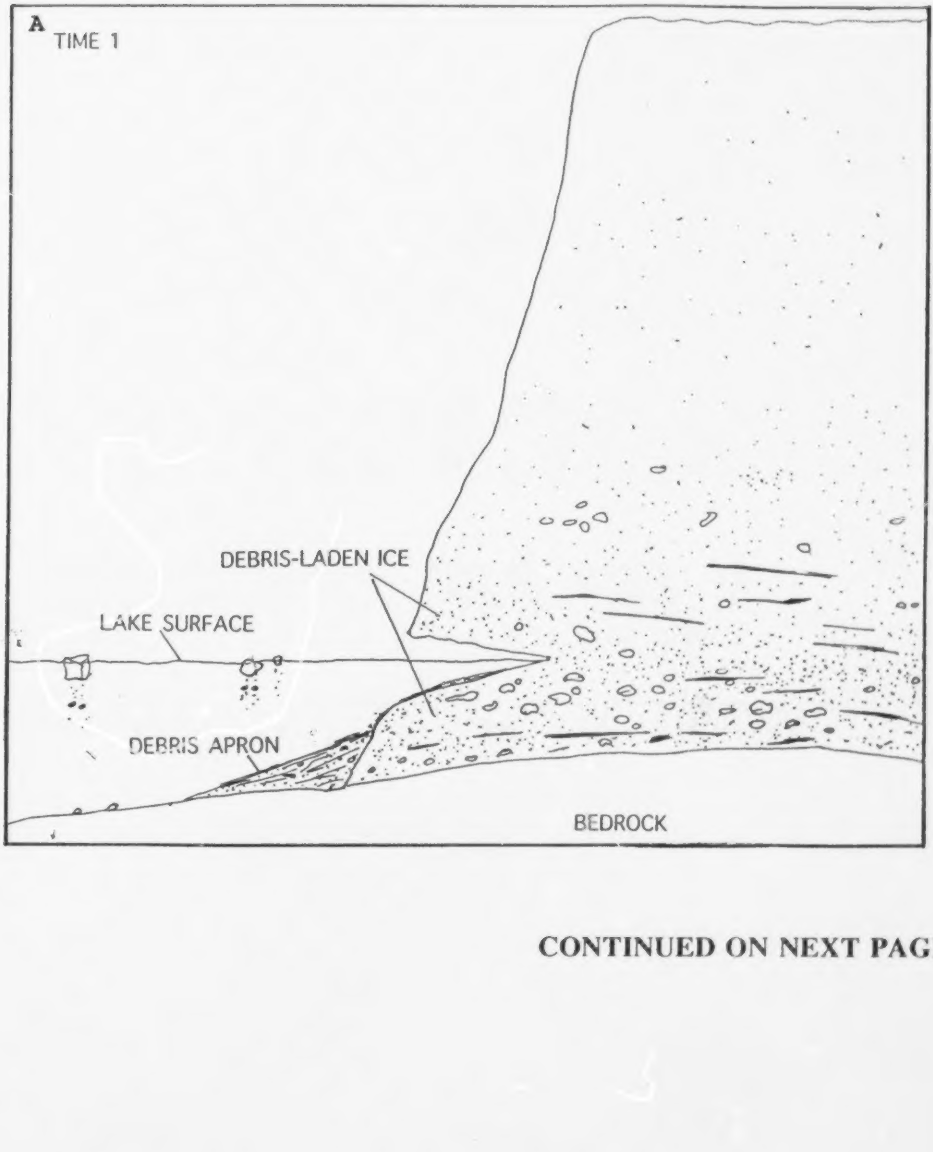
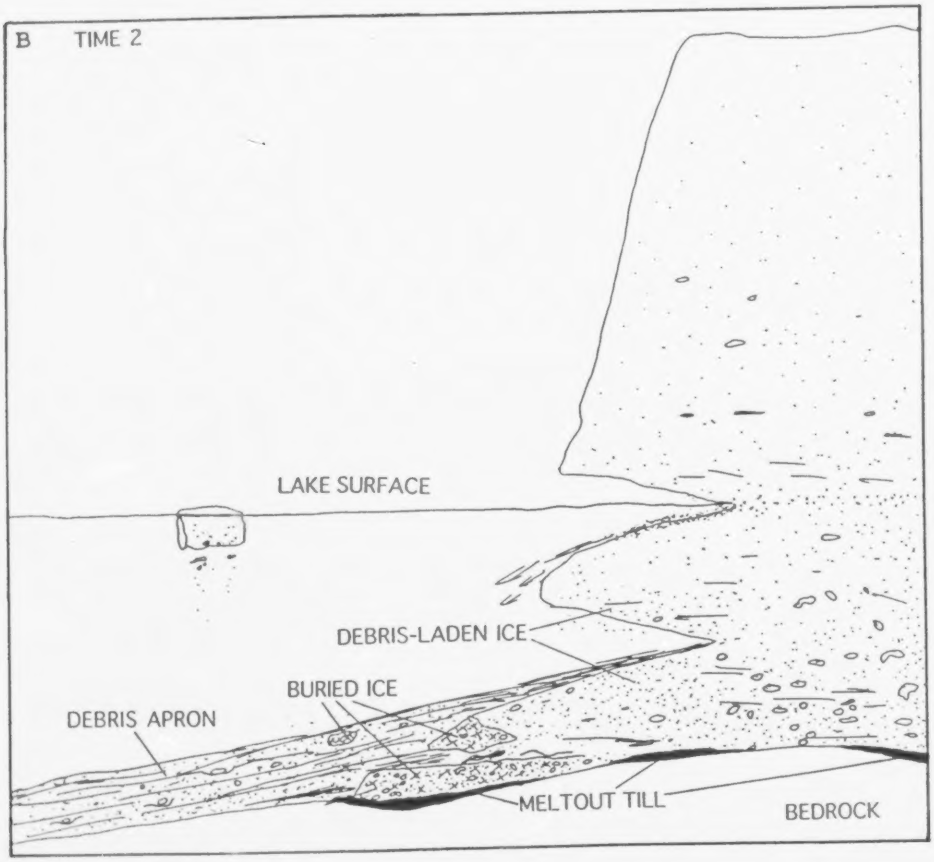

48
49
50
104
106
108
112
ALBERTA
SASKATCHEWAN
MONTANA
MH
SC
MC
OM
BP
WM
FC
MKR
MR
1
2
3
km
0
100

Figure 3.16 A, B. Schematic diagrams showing the development of a subglacial meltout till beneath a glaciolacustrine diamicton sequence. Time 1 during the early stages of lake infilling illustrates the initial formation of the ice apron and notch as well as development of debris apron in front of the ice. Time 2 after further infilling, illustrates the development of a second notch and the enlargement of the underlying ice apron and expansion of debris-covered area. Continued wastage of the ice where it contacts the lake would create an extensive buried ice apron. The lake could even become partially supraglacial. The depth of the lake can vary from a few metres to tens of metres. The buried apron may be a few hundred metres to several kilometres in extent.


A
TIME 1
DEBRIS-LADEN ICE
LAKE SURFACE
DEBRIS APRON
BEDROCK

CONTINUED ON NEXT PAGE


B
TIME 2
LAKE SURFACE
DEBRIS-LADEN ICE
BURIED ICE
DEBRIS APRON
MELTOUT TILL
BEDROCK

References cited.

Ashley, G.M., 1975, Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut, *in*: Jopling, A.V. and MacDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation, p. 304-320. Tulsa OK, S.E.P.M. Special Publication 23.

Ashley, G.M., 1989, Classification of glaciolacustrine sediments, *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam, p. 243-260.

Banerjee, I.,1973, Part A: Sedimentology of Pleistocene glacial varves in Ontario, Canada: Geological Society of America Bulletin, v. 226, p. 1-44.

Barnett, D.M. and Holdsworth, G., 1974, Origin, morphology and chronology of sublacustrine moraines, Generator Lake, Baffin Island, N.W.T., Canada: Canadian Journal of Earth Sciences, v. 11, p. 380-408.

Boulton, G.S., 1967, The development of a complex supraglacial moraine complex at the margin of the Sorbreen, Ny Friesland, Vestspitsbergen: Journal of Glaciology, v. 6, p. 717-736.

Boulton, G.S., 1968, Flow tills and related deposits on some Vestspitsbergen glaciers: Journal of Glaciology, v. 7, p. 391-412.

Boulton, G.S., 1975, Processes and patterns of subglacial sedimentation: a theoretical approach, *in* Wright, H.E. and Moseley, J.T., eds., Ice Ages: Ancient and Modern. Seel House Press, Liverpool. p. 7-43.

Brewer, R., 1976, Fabric and mineral analysis of soils, 2nd edition. Wiley, New York, 482 pp.

Bull, W.B., 1964, Alluvial fans and near surface subsidence, Fresno County, California. United States Geological Survey Professional Paper 437-A, 48 P.

Christiansen, E.A., 1968, Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 5, p. 1167-1173.

Christiansen, E.A. 1979, The Wisconsinan glaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Christiansen, E.A., and Sauer, E.K. 1988, Age of the Frenchman Valley and associated drift south of the Cypress Hills, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 25, p. 1703-1708.

Clayton, Lee and Moran, S.R. 1982, Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55-82.

Domack, E.W., and Lawson, D.E., 1985, Pebble fabric in an ice-rafted diamicton: Journal of Geology, v. 93, p. 577-591.

Dowdeswell, J.A., and Sharp, M.J., 1986, Characterization of pebble fabrics in modern terrestrial glacigenic sediments: Sedimentology, v. 33, p. 699-710.

Dreimanis, A.E., 1982, Two origins of the stratified Catfish Creek Till at Plum Point, Ontario: Boreas, v. 11, p. 173-180.

Dreimanis, A.E., 1989, Tills: their genetic terminology and classification, *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam p. 17-84.

Dyke, A.S., and Prest, V. K., 1987a, Late Wisconsinan and Holocene history of the Laurentide ice sheet: Geographie Physique et Quaternaire, v. 41, p. 237-263.

Dyke, A.S., and Prest, V. K., 1987b, The Late Wisconsinan Maximum; Geological Survey of Canada map 1702a, scale 1:5,000,000.

Evenson, E.B, Dreimanis, A., and Newsome, J.W., 1977, Subaquatic flow till: a new interpretation of the genesis of some laminated till deposits; Boreas, v. 6, p. 115-130.

Eyles, N., 1979, Facies of supraglacial sedimentation on Icelandic and Alpine temperate glaciers: Canadian Journal of Earth Sciences, v. 16, p. 1341-1361.

Feltham, K., 1989, Stratigraphy and distribution of Quaternary sediments in central Edmonton. Unpublished Masters Thesis, University of Alberta, Edmonton, Alberta, 232 p.

Frakes, L.A. and Crowell, J.C., 1967, Facies and paleogeography of Late Paleozoic diamictite, Falkland Islands: Bulletin of the Geological Society of America, v. 78, p. 37-58.

Fullerton, D.S. and Colton, R.B., 1986, Stratigraphy and correlation of the glacial deposits on the Montana Plains, *in* Richmond, G.M., and Fullerton, D.S., eds., Quaternary Glaciations of the United States. Quaternary Science Reviews, v. 5, p. 69-82.

Ghibaudo, G., 1992, Subaqueous sediment gravity flow deposits: practical criteria for their field description and classification: Sedimentology, v. 39, p. 423-454.

Gibbard, P., 1980, The origin of the Catfish Creek Till by basal melting: Boreas, v. 9, p. 71-85.

Gilbert, R., 1971, Observations on ice dammed Summit Lake, B.C., Canada: Journal of Glaciology, v. 51, p. 351-356.

Gilbert, R., and Shaw, J., 1981, Sedimentation in proglacial Sunwapta Lake, Alberta. Canadian Journal of Earth Sciences, v. 18, p. 81-93.

Griggs, G.B., and Kulm, L.D., 1969, Glacial marine sediments from the northeast Pacific: Journal of Sedimentary Petrology, v. 39, p. 1142-1148.

Gustavson, T.C., 1975, Bathymetry and sediment distribution in proglacial Malaspina Lake, Alaska: Journal of Sedimentary Petrology, v. 45, p. 450-461.

Hampton, M., 1972, The role of subaqueous debris flow in generating turbidity currents: Journal of Sedimentary Petrology, v. 42, p. 775-793.

Holdsworth, G., 1973a, Ice deformation and moraine formation at the margin of an ice cap adjacent to a proglacial lake *in* Fahey, B.D. and Thompson, R.D., eds., Research in polar and alpine geomorphology, 3rd Guelph Symposium on Geomorphology. Geo Abstracts, Norwich England, p. 187-199.

Holdsworth, G., 1973b, Ice calving into proglacial Generator Lake, Baffin Island, N.W.T., Canada: Journal of Glaciology, v. 112, p. 325-350.

Hooke, R.LeB, 1970, Morphology of the ice sheet near Thule, Greenland: Journal of Glaciology, v. 9, p. 303-325.

Howarth, P. J., 1968, A supra-glacial extension of an ice dammed lake, Tunsbergdalsbreen, Norway: Journal of Glaciology, v. 51, p. 413-419.

Klassen, R.W., 1989, Quaternary geology of the southern Canadian Interior Plains, p. 97-174. *in* Fulton, R.J., ed., Chapter 2 of the Quaternary Geology of Canada and Greenland, Geological Survey of Canada, Geology of Canada, v. 1. 839p.

Klassen, R.W., 1991, Surficial geology and drift thickness, Cypress Lake (72F) Saskatchewan: Geological Survey of Canada, Map 1766A (1:250,000).

Klassen, R.W., 1992, Nature, origin, and age relationships of landscape complexes in southwestern Saskatchewan: Geographie Physique et Quaternaire, v. 46, p. 361-368.

Kulig, J.J., 1985, A sedimentation model for the deposition of glacigenic deposits in central Alberta. [M.S. thesis], Edmonton, Alberta, Canada, University of Alberta, 245 p.

Kulig J.J., 1989, The glacial deposits within the Clover Bar Gravel Pit, Stop 29. in CANQUA 1989, Late glacial and post-glacial processes and environments in montane and adjacent areas field excursion guide p. 110-114.

Lambert, A., and Hsu, K.J., 1979, Non-annual cycles of varve-like sedimentation in Walensee, Switzerland: Sedimentology, v. 26, p. 453-461.

Lawson, D.E., 1979a, A sedimentological analysis of the western terminus of the Matanuska Glacier, Alaska: U.S. Army Cold Regions Research and Engineering Laboratory. Report 79-9. 122 pp.

Lawson, D.E., 1979b: A comparison of the pebble orientations in ice and deposits of the Matanuska Glacier, Alaska: Journal of Geology. v. 87, p. 629-645.

Lawson, D.E., 1989, Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits, *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam, p. 147-170.

Leckie, D.A., and Cheel, R.J., 1989, The Cypress Hills Formation (Upper Eocene to Miocene): A semi-arid braidplain deposit resulting from intrusive uplift: Canadian Journal of Earth Sciences, v. 26, p. 1918-1931.

Lindsay, J.F., 1966, Observations on the level of a self-draining lake on the Casement Glacier, Alaska: Journal of Glaciology, v. 45, p. 443-445.

Liverman, D.G.E., 1981, Sedimentology and drainage history of a glacier dammed lake, St. Elias Mountains, Yukon Territory. [M.S. thesis], Edmonton Alberta, University of Alberta, 221 pp.

Liverman, D.G.E., 1989, Sedimentology and history of a Late Wisconsinan glacial lake, Grande Prairie, Alberta, Canada: Boreas, v. 20, p. 241-257.

Mark, D.M., 1973, Analysis of axial orientation data, including till fabrics: Geological Society of America Bulletin, v. 84, p. 1369-1374.

Mark, D.M., 1974, On the interpretation of till fabrics: Geology. v. 8, p. 101-104.

May, R.W., 1977, Facies model for sedimentation in the glaciolacustrine environment: Boreas, v. 6, p. 175-180.

Morgenstern, R.C., 1967, Submarine slumping and the initiation of turbidity currents, *in* Richards, A.F., ed. Marine Geotechnique. University of Illinois Press, Urbana, p. 189-220.

Odynsky. W., and Newton, J.D., 1950, Soil Survey of the Wycroft and Watino sheets: Alberta Soil Survey Report 15, 84 pp

Odynsky, W., Wynn, K.A., and Newton, J.D., 1952, Soil survey of the High Prairie and McClennan Sheets: Alberta Soil Survey Report 17, 112p.

Ovenshine, A.T., 1970, Observations of iceberg rafting in Glacier Bay, Alaska and identification of ice-rafted deposits; Geological Society of America Bulletin, v. 81, p. 891-894.

Paul, M.A., and Eyles, N., 1990, Constraints on the preservation of diamict facies (melt-out tills) at the margins of stagnant glaciers: Quaternary Science Reviews, v. 9, p. 51-69.

Pierson, T.C., 1980, Erosion and deposition by debris-flows at Mount Thomas, New Zealand. Earth Surface Processes, v. 5, p. 227-247.

Pierson, T.C., 1981, Dominant particle support mechanisms in debris-flows at Mount Thomas, New Zealand and implications for flow mobility, Sedimentology, v. 28, p. 49-60.

Proudfoot, D.N., 1985, A lithostratigraphic and genetic study of Quaternary sediments in the vicinity of Medicine Hat, Alberta. [Ph.D. thesis], Edmonton, Alberta, University of Alberta, .

Ronnert, L. 1989, Basal till genesis in relation to bedrock morphology in the Berghem area, southwestern Sweden: Boreas, v. 18, p. 135-144.

Schmoll, H.R., 1961, Orientation of phenoclasts in laminated glaciomarine deposits, Copper River Basin, Alaska. United States Geological Survey Professional Paper 424c, article 219, p C192-C195.

Shaw, J., 1979, Genesis of Sveg tills and Rogen moraines of central Sweden: a model of basal melt out. Boreas, v. 6, p. 409-926.

Shaw, J., 1982, Meltout till in the Edmonton area, Alberta, Canada: Canadian Journal of Earth Sciences, v. 19, p. 1548-1569.

Shaw, J., 1987, Glacial sedimentary processes and environmental reconstruction based on lithofacies: Sedimentology, v. 34, p. 103-116.

Smith, D., and Ashley, G., M., 1985, Chapter 4, Proglacial lacustrine environment, *in* Ashley, G. M., Shaw, J., and Smith, N., D., eds., Glacial Sedimentary Environments. Special Publication 16, Society of Economic Paleontologists and Mineralogists, Tulsa, OK, p. 175-216.

Stone, K.H., 1963, Alaskan ice-dammed lakes: Annals of the Association of American Geographers, v. 53, p. 332-349.

Stringham, G.E., Simons, D.B.; and Guy, H.P., 1969, The behavior of large particles falling in quiescent liquids: United States Geological Survey Professional Paper, 562-C, 36p.

Sturm, M., and Matter, A., 1978, Turbidites and varves in Lake Brienz, Switzerland: deposition of clastic detritus by density currents. In Matter, A., and Tucker, M.E., (eds), Modern and Ancient Lake Sediments. Special Publication of the International Association of Sedimentologists. v2. Blackwell Scientific Publications, Oxford, p. 147-168.

Sugden, D.E., and John, B.S., 1976, Glaciers and Landscapes, A Geomorphological Approach. Edward Arnold Publishers, London. 376 pp.

Thomas, G.S.P., and Connel, R.J., 1985, Iceberg drop, dump and grounding structures from Pleistocene glaciolacustrine sediments, Scotland: Journal of Sedimentary Petrology, v. 55, p. 243-249.

Von Hoff, J.A., 1969, Tertiary gravels and sands in the Canadian Great Plains. Ph.D. thesis, University of Saskatchewan, Saskatoon, Saskatchewan.

Walker, R.G., 1984, General introduction: Facies, facies sequences and facies models, *in* Walker, R.G., ed., Facies Models (2nd. ed.), Geoscience Canada, Reprint Series 1, p. 1-11.

Westgate, J.A., 1968, Surficial geology of the Foremost - Cypress Hills area, Alberta: Research Council of Alberta Bulletin 22, 122 pp.

Westgate, J.A., 1972, The Cypress Hills. In Quaternary geology and Geomorphology between Winnipeg and the Rocky Mountains. International Geological Congress, 24th Session Montreal Quebec, Guidebook for Excursion C-22 p. 50-62.

Chapter 4: The origin of the Frenchman channel by the catastrophic drainage of lakes located on or near the southern flanks of the Cypress Hills.

A version of this paper will be submitted for publication in the Geological Society of America Bulletin.

Introduction

The Frenchman channel, the most prominent Late Wisconsinan landform on the Cypress Lake map sheet, is a sidehill channel that extends along the southern flanks of the Cypress Hills and Wood Mountain Uplands and terminates at the Milk River in Montana (Fig. 4.1). Radiocarbon dates between 3,000 and 11,000 BP obtained from wood contained in basal sediment from the Frenchman channel led Christiansen and Sauer (1988) to conclude that the channel was a Late Wisconsinan feature. Previous reconstructions (Christiansen 1979; Clayton and Moran 1982) had depicted the Frenchman channel as an interlobate channel, gradually eroded by the meltwater draining along it. A recent hypothesis (Vreeken 1991) suggests that the Frenchman channel south of Eastend was incised by subglacial meltwater flow. Shaw (1993, 1994, pers. comm.) theorized that the Frenchman channel was incised by subglacial megaflood described in Rains et al. (1993).

An alternative hypothesis that the Frenchman channel was not formed by gradual meltwater incision, nor by subglacial megafloods, but by the catastrophic release of meltwater impounded in glacial lakes on or near the Cypress Hills will be presented here. Focus will be centered on the formation of the upper segment of the channel, from Merryflat to 20 km south of Eastend, Saskatchewan (Fig. 4.1, 4.2, 4.3). This portion of the channel has been divided into four reaches with similar cross-sectional profiles and features (Fig. 4.2, 4.3). The implications of catastrophic, subaerial outbursts for the formation of the landscapes of central and southern Alberta, and to the deglaciation history of the area and the Prairies is also examined.

Location and geologic context

The Cypress Hills of southeastern Alberta and southwestern Saskatchewan form a preglacial plateau that rises up to 430 m above the prairie surface (maximum elevation 1410 m). The plateau consists of three blocks: the West Block, which extends 70 km into Alberta, and the Center and the East blocks, located entirely in southwest Saskatchewan (Fig. 3.1A). The West and Center blocks have steep northern margins and more gently sloping southern flanks, whereas the East Block on its eastern and northern flanks, rises gently from the prairie surface. The Frenchman channel extends 340 km, from its head at Merryflat, to its terminus at the Milk River valley in Montana (Figs. 4.1, 4.2, 4.3). This paper will discuss the formation of the channel from Merryflat to about 20 km south of Eastend.

Understanding the origin of the Frenchman channel requires a brief examination of the regional glacial history. At the Late Wisconsinan maximum the Cypress Hills were surrounded by the West and East lobes and crossed by the Gap Lobe (Fig. 4.4), (Klassen 1992; this study). The coalescence of the ice lobes impounded glacial lakes on and near the southern flanks of the Cypress Hills (Fig. 4.5; Chapter 3, this volume). This inferred ice-lobe coalescence differs from ice lobe patterns depicted in earlier reconstructions, in which the Frenchman channel was interpreted as an interlobate meltwater conduit between a more extensive West Lobe and a less extensive East Lobe (Christiansen 1979; Clayton and Moran 1982; Dyke and Prest 1987a). The glacial map of Canada (Dyke and Prest 1987b) also differs in that it shows the Gap Lobe as extending into Montana.

The ice lobe pattern suggested here (Fig. 4.4) has the East Lobe extending westward over the Frenchman channel onto the south flanks of the East and Center blocks and onto the Boundary and Old Man On His Back plateaus. The East Lobe impounded Lake Belanger between the Center and East blocks (Fig. 4.5). At the same time, the West Lobe was confined to the plains south of the Old Man On His Back and Boundary plateaus. The West Lobe was restricted to the plain southwest of Old Man On His Back Plateau because western-derived erratics are only found there and not on the East or Center Blocks, or northeast of the Old Man On His Back and Boundary plateaus (Klassen, pers com. 1989, 1990; Klassen 1989). Formation of the Frenchman channel in a sidehill position, cross-cutting the regional slope and bedrock obstacles requires that ice to the south.. If the East and West lobes had not coalesced, meltwater drainage would have followed the regional slope southeastwards along the flanks of the Old Man On His Back Plateau between the East and West lobes. The new ice lobe configuration means that the Frenchman channel was not an interlobate channel. A new explanation for its origin is therefore needed. In the following sections it will be shown that the Frenchman channel is a spillway, cut by the catastrophic drainage of glacial lakes impounded on or around the south slopes of the Cypress Hills.

Fieldwork

Sections containing glacigenic deposits are located on or near the flanks of the Cypress Hills. Three of these (the Canal, Gilchrist, and Cypress Lake sections described in Chapter 3) contain glaciolacustrine diamicton and rhythmites. Gravel pits, terraces, meltwater channels, and ice-pushed ridges, identified on air photographs, were also examined in the field for information pertinent to the formation of the channel (Fig. 4.2). The Frenchman channel itself, was also examined, in the field and on air photographs for evidence of its genesis. Finally, the elevational limit of granite, gneiss, and carbonate erratics on the southeast corner of the West Block was investigated.

Spillways, their formation and diagnostic features

Kehew and Lord (1986, p. 162) stated that spillways are the outlets of proglacial lakes and that these narrow, deep, trench-like features were created primarily by the rapid, extremely erosive fluvial activity characteristic of catastrophic glacial-lake outbursts. Spillways begin abruptly at glacial-lake outlets and generally end in other lake basins. They often cut across regional slope and may cross drainage divides and bedrock obstacles. Spillways have two major components: a broad outer zone and a deeply incised inner channel (Kehew and Lord 1986, 1987; Lord and Kehew 1987), (Fig. 4.6). The outer zone forms during the initial stages of release before a single dominant channel is incised. This outer zone may contain longitudinal grooves parallel to flow, shallow anastomosing channels, boulder lags, and streamlined erosional hills. The limits of the outer zone are often indistinct, because much of the flow was over ice (Kehew and Lord 1986,1987).

The inner channel is the most visible and diagnostic part of the spillway system and is believed to form in response to increased flow resistance following the formation of a boulder lag over the outer zone (Kehew and Lord 1986). The inner channel is trench-shaped (similar widths at the top and bottom of the channel), ranges from 25 to 100 m deep on average and has a high width-to-depth ratio. Non-catastrophic channels cut gradually by flowing water are asymmetrical and contain numerous slip-off slopes. The inner channels of spillways however, are symmetrical and lack well-developed slip-off slopes. In addition, the inner channel has regular meander bends (Clayton 1983) and may contain streamlined erosional remnants or infrequent bars of coarse sediment. The channels may have been used during several flood episodes and may have subsequently carried typical glacial meltwater streams. Today, spillways commonly contain underfit streams.

The large discharges carried by spillways had sufficient competence to transport the entire sediment load they contained through the spillway and deposit it in downflow glacial lakes. Large-scale point bars were occasionally deposited on the downstream side of meander bends or in the lee of landslides that had failed during incision of the spillway. Giant bars up to 2 km in length, 0.5 km in width, with sediment 20 m thick can be deposited. These bars contain unbedded matrix-supported very poorly sorted cobble gravel. Boulders up to 1 m in diameter are common but some are up to 3 m in diameter. Unbedded matrix-supported gravel may be deposited where the flow expands (Kehew and Lord 1987).

Rabot (1905) described the effects of outbursts from ice-dammed lakes in the Alps, Norway, Iceland, Greenland, and Alaska. Bretz (1923, 1981) was the first to suggest that landforms and deposits in the Channeled Scablands area of Washington were formed by catastrophic failure of ice dams impounding large ice-contact lakes. His work was followed by Pardee (1942) who described giant ripples, denuded bedrock benches and elevated eddy deposits emplaced during drainage of Lake Missoula. The first outburst channel described in detail on the Great Plains was the Souris spillway (Kehew, 1982). Additional spillway examples have been identified in many areas (southern Lake Agassiz: Clayton 1983; Kehew and Clayton 1983; Matsch 1983; Teller and Thorliefson 1983), (Kansas: Aber 1991), (southern Alberta: Evans and Campbell 1992), (Altay Mountains, Siberia: Baker et al. 1993), (Norway: Elfstrom 1987), (Lake Bonneville: Jarret and Malde, 1987), (Iceland: Maizels, 1989), (Grand Valley, Michigan, Kehew 1993). A controversial hypothesis describing a subglacial megaflood origin to large-scale landform assemblages has been advanced by Shaw and Kvill (1984), Shaw and Sharpe (1987), Shaw et al. (1989), Kor et al. (1991), and Rains et al. (1993). Shaw (1993, pers. com.) linked these subglacial megafloods to the formation of the Frenchman channel.

Description of the Frenchman channel

The upper Frenchman channel begins at a breach in the moraine (Fig. 4.2) crossing Merryflat and extends easterly along the south edge of the Cypress Hills to

Eastend where it turns to the southeast (Fig. 4.1, 4.2). It is mainly a sidehill channel that cuts across the regional slope. In places, it has cut into the Center and East blocks where they extend further to the south. In the portion of the channel investigated for this paper (from its head at Merryflat to about 20 km south of Eastend), (Fig. 4.1-4.3) the width and depth of the channel vary markedly. Slip-off slopes, terraces, and other indications of gradual incision are absent. Broad benches, covered with boulder lags containing scattered erratics, flank parts of the channel (Feature 5, Fig. 4.2). The channel is predominantly straight, or nearly straight, with a few uniform meander bends (Fig. 4.3). The stretch of channel studied was divided into four reaches (Fig. 4.1). Within each reach the cross-sectional area remains relatively constant and the morphological features are similar. Reach 1 extends from Merryflat to Cypress Lake. Reach 2 from Cypress Lake to Palisades Coulee. Reach 3 from Palisades Coulee to Eastend and Reach 4 from Eastend to about 20 km south. In addition to these four reaches, the Palisades and Eastend coulees that connect to the Frenchman channel in Reach 3 and 4 respectively, are described since they also had a role in forming the upper reaches of the Frenchman channel. A series of profiles for the Frenchman channel and the Palisades and Eastend Coulees (Figs. 4.3, 4.7, 4.8) illustrates the cross-sectional variability.

Reach 1. Merryflat to Cypress Lake.

The breach in the moraine (Fig. 4.2, Features 2 and 3 respectively, Fig. 4.9) crossing Merryflat marks the abrupt beginning of the channel (Fig. 4.7 Profile 1). The breach is 0.5 km wide and 23 m deep and is incised into the gravel of the flats. A set of channels, also incised into the gravel of Merryflat, extends northwest to southeast across the flats and coalesces at the entrance to the moraine breach (Feature 2, Fig. 4.2). The longest of the channels extends onto the West Block and is cross-cut by Battle Creek (Fig. 4.2). On the interfluve between Battle and Adams creeks is a set of channel scours that drop progressively in elevation (Fig. 4.2, Feature 1). Immediately south of the breach the channel increases in width to about 0.75 km but maintains its depth of about 25 m (Profile 2, Fig 4.7).

About 3 km south of Profile 2 the channel turns east and extends along the southern flanks of the Center Block. Here the channel is obscured by infilling during a subsequent Late Wisconsinan readvance and erosion by Battle Creek (Profile 3). At Profile 4, the shape of the channel is much more pronounced. The channel is about 1 km wide and 40 m deep. A broad bench at 1070 m elevation, 95 m above the base of the profile, is evident. Profile 5, across the west arm of Cypress Lake, is more irregular. The trench-like channel is still evident and a broad bench occurs at about 1190 m elevation. These benches lack fine grained glacial diamicton but erratics are scattered across their surfaces. Quartzite cobbles and boulders from the Cypress Hills Formation make up the bulk of the material on the bench tops.

Reach 2. Cypress Lake to Palisades Coulee.

The profile at Cypress Lake (Profile 6, Fig. 4.7) indicates that the channel has widened considerably. The almost complete absence of a south wall is particularly evident (Fig. 4.10). The north wall is about 75 m high, but the poorly developed south wall, where present, is only 8 to 15 m high, of which 5 to 8 m is glaciolacustrine diamicton (Chapter 3, this volume) deposited in a glacial Lake Cypress. Klassen (1991, 1992) mapped extensive hummocky diamicton and glaciolacustrine sediment in this area.

Also on the north side of the channel in this area a meltwater channel system extends along the eastern edge of the Gap Lobe moraine. At the southern terminus of the moraine (Feature 3, Fig. 4.2), the channel turns eastward along the north bank of the Frenchman channel, where it becomes discontinuous and difficult to trace.

East of Cypress Lake, where the channel cuts through a bedrock obstruction, the channel narrows to approximately 800 m width at its base and is about 45 m deep. It is surrounded on both sides by broad flat benches that lack diamicton and are covered with a lag of boulders and cobbles from the Cypress Hills Formation along with scattered erratic boulders (Fig. 4.2, Feature 5), (Fig. 4.11, 4.12).

At Profile 8 the channel is narrower (about 0.7 km at its base), deepens to about 75m and has a broad, gently sloping bench on its north side. An unusual set of streamlined bedrock knobs (Fig. 4.2, Feature 6) is present in the channel just before the junction with Palisades Coulee. The knobs occur from near the rim of the channel to its base (Fig. 4.13). The pre-Pleistocene formations within the knobs are horizontal and in their original stratigraphic sequence.

Reach 3. Palisades Coulee to Eastend.

Profiles 9, 10, and 11 illustrate the dramatic change in the Frenchman channel from 3 km west of Ravenscrag to the Eastend (Fig. 4.14). In this reach, the channel is 1.0 to 1.5 km wide and 115 to 145 m deep. Rotational slump blocks, that collapsed into the channel, have reduced its width and depth and impart a notched profile to the channel from Ravenscrag to Eastend. Drilling by Christiansen and Sauer (1988) revealed from

50 and 80 m of colluvial and alluvial fill within the channel between Cypress Lake and Eastend.

The Palisades Coulee (Profiles 12, 13, 14, Fig. 4.8) joins the Frenchman channel three kilometres west of Ravenscrag. It is at this junction that the cross-sectional area of the Frenchman channel increases markedly (Profiles 9-4, Fig. 4.7). Above 990 m elevation, the coulee has a deeply incised, sharp-sided, trench-like, cross-sectional profile and cuts across the regional slope to join the Frenchman channel. Below 990 m elevation the profile is asymmetric with a terrace at about 990 m. On the east side of the coulee a series of shallow northeasterly trending gullies terminates in the Frenchman channel (Fig. 4.2, Feature 7). Two braided esker systems feed into the bifurcated head of the Palisades Coulee (Fig. 4.2, Feature 8).

Reach 4. Eastend and southward

At Eastend the Frenchman channel turns south and extends across the Frenchman Plain. Here the channel is joined by Eastend Coulee which extends along the eastern edge of the East Block and terminates at Eastend. Profiles 15-17 (Fig. 4.8) illustrate the shape of the Eastend Coulee.

Also at Eastend, on the inside bend of the south wall, are nested scour channels (Fig. 4.2, Feature 9). Gravel pits in the scour channels contain sand, gravel, and large boulders of up to 3 m^3.

Profiles 18, 19, and 20 (Fig. 4.7) show the Frenchman channel south of Eastend. The channel is no longer as deeply incised but its trench-like cross-sectional profile is maintained (Fig. 4.15). Width increases from about 1.25 km at P18 to 1.5 km at P19 and reaches a maximum of about 2 km at P20. Depth also increases southwards. At P18 it is about 55m, at P19, 70 m, while at P20 it is between 70 and 85 m. At P18 broad flat benches lie on both sides of the channel. At P19, the terrain to the east slopes gently into the channel, but the surface to the west slopes away from the channel. At P20, broad benches on both sides slope gently into the channel.

Formation of the Frenchman channel

The symmetrical trench-like cross-sectional profile, cross-topographic trend, incision through bedrock obstructions and lack of slipoff slopes, and terraces, suggest that the Frenchman channel is a catastrophically incised spillway similar to those described by Kehew and Lord (1986, 1987). There is a lack of glaciofluvial stream sediment within the channel. The perched gravel pits at Eastend are on the inside of a large meander bend where the channel turns to the southeast. Kehew and Lord (1987) noted that outburst gravels are preferentially found on the inside of bends in spillways. The location of the Eastend gravel pits agrees with a catastrophic origin for the channel but normal fluvial erosion would also deposit gravel in this location. But the boulders in the gravel pits at Eastend are much larger than those associated with normal fluvial flow rates. Additionally there are no gravel terraces that connect to the Eastend pits. They exist in isolation.

The broad gravel and erratic-covered benches, that surround the channel from just west of Cypress Lake to near the junction with Palisades Coulee, correspond to the broad outer zone formed before the main channel became incised (see Kehew and Lord 1986, 1987). The outer zone forms before an inner channel can be incised. The outer zone may also carry meltwater that could not be accommodated within the inner channel itself. Their disappearance where the channel becomes deeply incised in Reach 3 (Fig. 4.14) indicates that the channel in this area was able to transport all the meltwater.

The Palisades and Eastend Coulees are also symmetrical and trench-like in profile. They do not contain stream sediments, or terraces and they begin abruptly. These features indicate that these coulees are spillways.

Kehew and Lord (1986, 1987) state that spillways maintain an approximately constant width and depth unless there is input from additional water sources. The marked change in west to east cross-sectional profiles reflects the materials that form the walls of the channel and inflow from several water sources along its length. Reach 1 is only 20 m deep and 0.5 km wide, with only the north wall continuously present. Christiansen and Sauer (1988) concluded that ice to the south was a major reason of the sidehill positioning of the Frenchman channel. Where the south wall is absent in Reach 1, it must have been formed by ice. Kehew and Lord (1986) noted that where all or part of the walls and base of a spillway were formed by ice, the spillway can lose its distinct profile and many of the associated features. Because ice would be more susceptible to meltwater erosion than the gravel and boulder-armoured flanks of the Cypress Hills, the channel here may have been much broader. Identifying the true width of the channel in Reach 1 is difficult because a readvance partially infilled the channel (Chapter 6, this volume).

In Reach 2, the channel abruptly broadens at Cypress Lake and the south wall of the channel is again absent (P10, Fig. 4.7, Fig. 4.10). Immediately east of the lake, the channel narrows as it passes through a bedrock obstruction. Broad benches are observed in the bedrock surrounding the channel there (P7, Fig 4.7, Fig 4.11, Fig. 4.12). The broadening of the channel occurs where the outburst entered Lake Cypress (Chapter 3, this volume). The absence of a south wall means that the outburst in this area must have been confined to the south by an ice wall. The narrowing of the main channel east of the lake reflects the change from ice-confined flow to bedrock-confined flow. Because the bedrock would be more resistant to lateral erosion than ice, the outburst could not spread laterally; instead a deep narrow trench with broad outer zone formed. The streamlined bedrock knobs at the end of Reach 2 (Fig. 4.13) may be similar to streamlined bedforms described by Clayton (1983) and Kehew and Lord (1986). Their preservation may reflect a flow of insufficient energy to remove them.

Reach 3 (145 m deep, 1.5 km wide, Fig. 4.14) is the most entrenched of the four reaches. The true width and depth are difficult to estimate as colluvial and alluvial sediment and rotational slumping have partially infilled the channel. Maximum channel entrenchment occurs immediately downstream from the junction with the Palisades Coulee. This suggests a large influx from the coulee. The entrenchment and lack of a broad outer zone here suggest that the outburst was totally contained within the confining bedrock walls.

The Palisades Coulee cuts across the topographic trend to join the Frenchman channel. Small channels (Fig. 4.2, Feature 7) extend across the east wall of the coulee before linking up with the Frenchman channel. These small side channels may be similar to offshoot channels (Kehew and Lord 1987) formed during the early stages of breakout before a single channel becomes the dominant path of the outburst.

The Palisades Coulee has a symmetrical trench-like profile near its top, but an asymmetric one at its base (Fig. 4.8, P12, 13, 14). Hummocky terrain surrounding the coulee may explain this departure from a normal spillway profile. With the incision of the Frenchman channel and formation of the coulee, ice surrounding the coulee was isolated and became stagnant. The deep coulee subsequently captured most of the drainage from the hummocky terrain as indicated by the large braided eskers (Fig. 4.2, Feature 8) observed in the channels joining from the hummocky terrain to the Palisades coulee (Fig. 4.2). This later meltwater flow eroded a shallow meandering path into the soft shale base of the coulee, creating the asymmetric profile below an elevation of 990 m.

The gravel pits at Eastend (Fig. 4.2, Feature 9) may contain the only direct evidence of sedimentation from the floodwaters. The large boulders (up to 2 by 1 by 1 m) in these pits are unlikely to have been transported by normal fluvial flow. The scour channels (Fig. 4.2, Feature 9) at several elevations may indicate rapid downcutting perhaps during several pulses.

South of Eastend the channel is less entrenched (Fig 4.7, P18, P19, P20, Fig. 4.15), and is surrounded by broad benches that gently slope into it. The outburst south of

Eastend may have followed a pre-existing supraglacial stream course, or may have followed the surface topography of the ice. Free of the confining bedrock walls, the outburst was able to expand laterally by thermally eroding its confining ice walls. The broad benches that slope gently into the channel on Profiles 18, 19 and 20 are outer channel areas. At Profile 19, the terrain slopes away from the west side of the channel. This indicates that either the flow was confined solely to the channel or that ice prevented any spillover to the west.

The change in the west to east cross-sectional profiles reflects two factors. First, where the channel is shallow and broad, it was confined to the south by ice that was readily eroded. Where the channel was confined by bedrock it is narrow and entrenched. Furthermore, the profile changes coincide with increased discharge. The largest water inflow appears to have entered from Palisades Coulee.

The channel has a sidehill position because it was confined by a large ice mass to the south, but its location along the flanks of the Cypress Hills also reflects the change in the character of the ice where it ascended onto the hills. During the early stages of deglaciation, meltwater was transported along the margins of the ice lobes in channels that were at least in part supraglacial, as shown by their present discontinuous nature. Initially the outburst followed similar supraglacial channels on the south flank. The outburst rapidly eroded through the thin ice on the south flanks of the Center Block to the underlying bedrock. Because it was easier to laterally erode the ice than to cut down into the bedrock, the flow shifted downslope faster than it became entrenched. Where the ice ascended onto the south flanks of the Center Block, the ice was stressed, crevassed, and weakened. When the outburst reached the relatively thick, stressed, crevassed ice along the inflection point, it cut rapidly downwards. The confining ice walls to the south were now thick enough to slow the lateral movement of the outburst, and the channel became fixed in a sidehill position. In Reach 3, the flow became even more restricted between the high bedrock walls. At Eastend the channel turns southward and is no longer in a sidehill position. The channel is broadest here. Initially the channel may have been following a pre-existing supraglacial meltwater channel or was controlled by the topography of the ice in the reach. Since the ice walls here confined the outburst less, it was able to spread laterally. The broad benches on either side of the channel formed before the main channel became established.

The features of the channel, the changes in cross-sectional profile, and its location all demonstrate that it is a spillway incised by the catastrophic release of meltwater. It has been suggested by Shaw (1994, pers. comm.) that the Frenchman channel is a tunnel valley cut by catastrophic sub-glacial drainage. The progressive increase in channel size from reach 1 to 4 indicates increased carrying capacity downstream. A channel cut by a non-local megaflood would be expected to have a fairly constant once input from local water sources is accounted for. Since many of the features of tunnel valleys and spillways are similar (see Shaw 1983 and Brennand and Shaw 1994). This is a possibility that cannot be ignored. The author agrees that in some cases subglacial catastrophic drainage was the initiating factor in forming a channel. The main difference in viewpoint is whether the drainage is from a locally impounded ice-contact lake or extra-local drainage. This point will be examined in more detail later. Possible source(s) of water for its incision will now be examined.

Sources of water for channel incision.

For the Frenchman channel to be a spillway it must have drained glacial lakes. The glacial lakes (Fig. 4.5) described in this section have all been named for creeks or geomorphic features near the location of the proposed lakes. Lake Graburn is named for Graburn Gap, on the north side of Merryflat. Lake Cypress is named for the present Cypress Lake, which occupies a small part of the basin of the former lake. Lake Belanger is named for Belanger Creek, which drains southwards from the site of this former lake. Lakes Fairwell and Blacktail are similarly named for creeks. Lake Robsart is named for the town of Robsart (Fig. 4.1, 4.2), near the site of the proposed lake.

Formation of the Frenchman channel began when drainage of Lake Graburn formed Reach 1. Reach 2 transported this water. To it was added the water released from the outburst of lakes Cypress and Belanger. In Reach 3, there is a marked increase in cross-sectional area of the channel at its junction with Palisades Coulee. This increase reflects the additional inflow from several lakes on the flanks of the East Block (lakes Blacktail and Fairwell on the south flank of the East Block and south of the channel, Lake Robsart). At Eastend a much later inflow from a lake north of the Cypress Hills incised Eastend Coulee and reused part of the Frenchman channel. No water sources have been identified south of Eastend but thermal erosion of the ice confining the outburst would have added additional water to the outburst at many locations along the channel.

Chapter 3 (this volume) identified two glacial lakes, Belanger and Cypress (Fig. 4.5), close to the channel. Lake Belanger formed when the East Lobe ascended onto the southern flanks of the East and Center blocks, impounding meltwater drainage, that flowed across the Center and East blocks from the northern ice mass. The lake contained between 6×10^9 and 1.0×10^{10} m^3 of water. The lake drained along Sucker and Davis creeks, that join the channel just east of Cypress Lake (Fig. 4.2, 4.3).

Lake Cypress formed, during the initial stages of retreat, in the coalescence area of the three ice lobes. This lake contained about 4 x 10^9 m^3 of water. Neither of these lakes contained enough water to cut a channel on the scale of the Frenchman channel and both of these lakes lay east of the head of the Frenchman channel. The initial reach of the Frenchman channel (from Merryflat to Cypress Lake) therefore was cut by water from a different source. Other lakes impounded on or around the southern flanks of the Cypress Hills catastrophically drained and contributed to the incision of the Frenchman channel to the east. The cumulative increase in discharge from the outburst of each successive lake caused the west to east enlargement of the cross-sectional profile of the channel.

It was the catastrophic release of Lake Graburn that triggered the subsequent release of the other lakes. The strongest evidence that the lakes burst in succession is the progressive increase in channel size from its head at Merryflat to south of Eastend. There is no evidence within the channel of multiple terrace levels that could be interpreted as multiple releases from other glacial lakes. But the nested scour channels at Merryflat and Eastend may indicate some type of pulsed flow.

Reach 1. Lake Graburn

Merryflat is a depression between the West and Center blocks (Fig. 4.2, 4.9), with a setting very similar to that where Lake Belanger formed. Before the advance of the West lobe, meltwater from the ice north of the Cypress Hills flowed across the Cypress Hills (Klassen 1992). The arrival of the West Lobe likely impounded this southward flowing meltwater, forming a lake. The moraine that extends across the flat (1095 m elevation for the moraine top), was deposited from this ice. The ice reached an elevation of about 1140 m elevation on the south flanks of the West Block. The ice therefore extended a minimum of 55 m above the top of the moraine. This ice barrier impounded a lake over the flat and onto the flanks of the West Block. Fresh unweathered erratics (2 to 3 m^3) overlie the preglacial gravel on the south slope of the West Block to an elevation of about 1250m. They are lie several kilometres beyond the limit of the Late Wisconsinan ice (the most extensive ice to affect the area) and were most likely ice-rafted across the impounded lake. This was Lake Graburn.

Westgate (1965a, b, 1968, 1972) concluded that these erratics, beyond his Late Wisconsinan limit, were evidence of an earlier more extensive glaciation. Vreeken (1986) discussed why these erratics were unlikely to have been deposited by an older advance and proposed that they were ice rafted across small ice-marginal ponds. Since there are no obstructions along the southern slope of the West Block, it is more likely that a single long narrow ice-marginal lake extended from Merryflat to the western edge of the West Block (Fig. 4.5). If the maximum elevation of the erratics corresponds to the surface of the lake, then the impounded lake would be 160 m deep. The minimum size of this lake would have been 40 km long, about 160 m deep and from 12.5 km wide in the Merryflat area to 5 km wide at its western terminus. It would have contained between 3 x 10^{10} and 7.5 x 10^{10} m^3 of water. This water volume should have been sufficient to cut the channel from Merryflat to Cypress Lake but investigations into non-local sources of water input in this area are planned.

The absence of glaciolacustrine sediment in the former lake basin means that either the lake was too short lived for sediment accumulation or that input of fine-grained sediment was low and did not form a thick sediment layer in the lake. Embleton and King (1975) observed that most of the fines were removed from glaciolacustrine lake basins during rapid drainage. Any unstable unconsolidated sediments deposited on the steep slopes of the West Block, could have been rapidly eroded following drainage of the lake. Glacial lake sediments deposited along the base of the West Block would be obscured by later slumps and colluvium.

The evidence for Lake Graburn is largely circumstantial, and the possibility that there is a water source outside the area as suggested by Shaw (pers. comm., 1993) must be considered. The two most likely water conduits are the Battle Creek channel and the Middle Creek channel (Fig. 4.2). The Battle Creek channel begins at Elkwater Lake and extends across the West Block. This channel is large and crosscuts the smaller channels that extend across Merryflat to the moraine breach that mark the start of the Frenchman channel. Since the Battle Creek channel crosscuts them, the Battle Creek channel must be younger. Furthermore, the Battle Creek channel can be traced southwards from Merryflat on air photographs (Fig. 4.2).

Inflow of water along Middle Creek channel is also unlikely. The Middle Creek channel is a product of a later readvance because when the Frenchman channel formed, the Middle Creek channel area was covered by ice (Chapter 6, this volume). The character of the Middle Creek channel is also different. The Middle Creek channel is small and contains abundant slipoff slopes and asymmetric valley profile of the Middle Creek channel are diagnostic of gradual fluvial processes, not catastrophic cutting and spillway formation. Finally the head of the Middle Creek channel in Alberta, is in a hanging valley position above the Medicine Lodge channel and could have carried only a small amount of spillover, if any, from the Medicine Lodge channel. The Middle Creek channel therefore was not a conduit for transporting the large volumes of water needed for cutting the Frenchman channel.

Reach 2 and 3. Lake Robsart

A large ice-contact lake was located in the area southwest of Ravenscrag (Fig. 4.2, 4.5). This lake was the most extensive in the system and straddles Reaches 2 and 3. It formed in the coalescence area between the East and West lobes and may have been partially supraglacial as indicated by the hummocky lake sediment and ice-walled lake plains south and south east of Cypress Lake and in the area north of the Old Man On His Back Plateau (Fig. 4.1, 4.2, 4.5), (Klassen and Vreeken 1987; Klassen 1991,1992). Excavation through the rim of one of these plateaus (Klassen and Vreeken 1987) exposed rhythmites over clayey diamicton deposits. The supraglacial aspect of this lake is unusual but it may have been similar to the ice-stagnation lake networks described in Ashley (1989) or the long lasting supraglacial ice-walled lakes described by Clayton and Cherry (1967). The infilling and outburst of Lake Cypress triggered the release of Lake Robsart. It is unknown how interconnected the two lakes were. The extent of Lake Robsart is known only from the distribution of lake sediment. Its depth and the volume of water it contained are also unknown. As the extent of this lake and its depth are uncertain at this time only a speculative reconstruction based on the distribution of ice-walled supraglacial lakes in the hummocky terrain and the hummocky lake sediment (Klassen 1992) is presented in Figure 4.5.

The main drainage passage for Lake Robsart was the Palisades Coulee in Reach 3. Drainage of the lake through Palisades Coulee explains the south to north drainage direction of the coulee, its trench-like appearance, and why the coulee cuts through a bedrock knob and across the regional slope before it joins the Frenchman channel (Fig. 4.2). A large inflow from this lake along Palisades Coulee would also explain the marked increase in the cross-sectional area of the Frenchman channel (Fig. 4.8, Profiles 10,11) at its junction with this coulee.

Reach 3. Lakes Blacktail and Fairwell

Glacial Lake Blacktail, the smallest of the lakes, was located on the lowest bench of the East Block in the Blacktail Creek area (Fig. 4.2, 4.5). Erratics in the area beyond the limit of the last ice have a fresh unweathered appearance and rest on preglacial gravel. They were probably transported by ice rafting across the glacial lake. A system of low relief channels draining only the bench top around Blacktail Creek and not connected to the cross-hill meltwater channel system that flowed over the East Block (Klassen and Vreeken 1987, Klassen 1992), (Fig. 4.2, 4.5) formed during its drainage.

Ice deposited a moraine at an elevation of 1140 m on the bedrock bench northwest of Fairwell Creek. It is possible that the cross-hill meltwater flow was impounded between the ice and the higher benches of the East Block, forming a lake that covered much of the southern flanks of the East Block. Perhaps even extending to Lake Belanger. This possibility requires further investigation.

Melting of confining ice as an additional water source.

The Frenchman channel was bounded by ice on one or both sides along much of its course. The thermal erosion of ice as the outburst was forming its channel would have supplied a significant amount of water to the system. The amount of meltwater from this source is impossible to determine exactly. If the outburst cut a channel south of Eastend into the ice that was 1.5 km wide and 200 m deep, 3.0×10^7 m^3 of meltwater would have been released for each kilometre of ice channel cut. Similar thermal erosion of ice along the length of the channel, such as at Cypress Lake, would have supplied a significant volume of water to the channel.

Summary

The evidence for glacial lakes on and around the southern flanks of the Cypress Hills is fragmentary. When glacial ice was present to the south, cross-hill meltwater flow from the northern ice would have been impounded in the depressions between the blocks that make up the Cypress Hills. Glaciolacustrine sediment has, so far, been found in the areas covered by lakes Belanger and Cypress (Chapter 3, this volume) and in ice-walled lake plains north of the Old Man On His Back Plateau near the Palisades Coulee (Klassen and Vreeken 1987). More investigation is needed in the hummocky terrain surrounding the Palisades Coulee to determine the origin of the hummocks and their relationship to the ice-walled lake plains there. Further research is also needed to clarify the number and distribution of lakes in the area and to evaluate the possibility of non-local water inflows.

Drainage history

The lakes, on and around the southern flanks of the Cypress Hills, supplied the water for cutting the Frenchman channel from Merryflat to Eastend. During the glacial maximum, three lakes (Belanger, Graburn, and Blacktail, and perhaps Fairwell) were present on the southern slopes and Lake Robsart was infilling on the plain. Lake Cypress had not yet formed. At this time, meltwater was transported in small channels, along the margins of the ice lobes (Fig, 4.2, Feature 4, Fig. 4.4, 4.5).

During the initial stages of retreat, Lake Cypress formed near the confluence of the three lobes. Ice dams continued to block lakes Graburn, Belanger, Robsart, Blacktail, and Fairwell(?). Meltwater continued to flow in the ice marginal and supraglacial channels.

Continued ice retreat and thermal erosion by meltwater led to thinning of the ice impounding Graburn Lake and ultimately the catastrophic release of the lake. This breached the moraine across Merryflat. The initial breakout followed the pre-existing ice marginal and supraglacial channels. The flat erratic-covered terraces north of Cypress Lake and near the Gilchrist section and Fairwell Creek (Fig. 4.4, Profiles 7 and 8) correspond to the outer channel formed during the early stages of breakout (Kehew and Lord 1986, 1987). The outflow rapidly shifted to the inflection point where the thick ice on the plain was stressed and fractured by its ascent onto the southern flanks of the Cypress Hills. The floodwater easily eroded the ice at the inflection point, and the inner channel of the spillway became established there. The outburst continued to Lake Cypress. The rapid inflow of water into Cypress Lake rapidly infilled the lake and forced the incision of a narrow channel across the bedrock obstruction east of Cypress Lake. This bedrock obstacle confined the flow and prevented it from altering its course. Cutting of the channel past Cypress Lake also triggered the release of lakes Belanger, Blacktail, and Fairwell(?) along Sucker, Blacktail, and Fairwell creeks respectively. The outburst also initiated the catastrophic release of Lake Robsart, through Palisades Coulee. The large inflow of water from the coulee deeply incised the channel. At Eastend the flow was no longer confined between bedrock walls and was able to spread laterally causing the channel to become broader and shallower. Pulses during the rapid downcutting eroded the scours on the inside bend of the channel at Eastend.

After incision the Frenchman channel, the hydrology of the area was permanently altered. The Palisades Coulee transported meltwater from the stagnant ice mass isolated south of the Frenchman. These meltwaters eroded a meandering channel system into its base. The Frenchman channel carried meltwaters from later readvances as well as drainage from Eastend Coulee. Presently, a misfit stream flows within it.

Discussion

Interpreting the Frenchman channel as a catastrophically cut spillway is a major revision of earlier ideas. The Frenchman channel is not a simple spillway that fits neatly into the model developed by Kehew and Lord (1986, 1987). It is a complex feature influenced by three ice lobes, several glacial lakes, transitions from an ice-walled to a bedrock-walled channel, and the presence of the Cypress Hills upland. Post-incision changes have greatly modified the cross-sectional profiles of the channel and the channel's features. The size of the channel reflects the fact that flow in Reaches 1, 2 and 4 were predominantly ice walled whereas in Reach 3, flow was confined between bedrock walls. At Merryflat, the Frenchman channel has been altered by the formation of the Battle Creek and Middle Creek channels and a subsequent advance (the Middle Creek Advance). The Battle Creek channel crosscuts the Frenchman channel modifying the head of the Frenchman channel. The Middle Creek Advance redirected meltwater discharges down the Frenchman channel, eroded older sediment and landforms, and deposited younger ones. Thus a very complex glacial history was compressed into a few square kilometres.

At Cypress Lake and Eastend, the confining ice, that once formed part of the walls of the channel, has melted. Numerous slump block failures between Ravenscrag and Eastend partially infilled the channel, obscuring its true width and depth. Inflow from Eastend Coulee into the Frenchman channel at Eastend has also changed the channel there.

Possible sedimentological evidence for the catastrophic drainage of Lake Belanger is seen in the top of the Canal section. Diamicton units composed of broken silt and clay rhythmites may have formed when lacustrine sediments slumped following rapid drainage of the lake. Liverman (1981) observed numerous small slope failures and rotational movements as the level of an ice-dammed glacial lake dropped during rapid drainage. The presence of several diamicton layers composed of broken silt and clay rhythmites in the upper part of the Canal section may indicate that Lake Belanger burst more than once.

The mechanism for catastrophically releasing the lakes is uncertain. Thorarinsson (1939) thought that drainage of ice-dammed lakes occurred when the water in the lake had sufficient volume to float the ice dam. Glen (1954) combined floating of the ice margin with the formation of a subglacial tunnel, enabling ice-dammed lakes to drain more fully. Stone (1963), Mathews (1965), Lindsay (1966), and Howarth (1968) observed both the floating of the ice margins and ice tunnel formation in ice-dammed lakes. Most of these ice-dammed lakes filled and drained several times. These mechanisms would likely have occurred during the initial stages of drainage. It is possible that the lakes in the study area filled and emptied several times. The evidence is presently too fragmentary to draw firm conclusions.

The source(s) of the water that incised the Frenchman channel requires additional investigation. If the channel south of Eastend attained bankful conditions, it could transport 1.3 x 10^5 m^3/s; the source of such a large volume of water is still only hypothetical. Except for lakes Belanger and Cypress, the evidence for the other lakes and their extents is fragmentary and circumstantial. The extent of the lake that burst to erode the Palisades Coulee requires further investigation as does the possibility that a large lake was impounded between the ice and the highest parts of the East Block. Other sources of water such as the melting of confining ice should also be explored.

The marked increase in channel cross-section from Ravenscrag to Eastend may not only reflect additional water inputs but the change from an ice-walled channel to a bedrock-walled channel. The effects of flow confinement in Reach 3 were probably substantial.

It is possible that the outburst exploited a pre-existing but shallower river valley, especially in Reach 3. An approximate elevation for an older channel may be established by identifying the formation that forms the tops of the slump blocks. If the flow in this reach was confined to an older channel, the tops of the slump blocks could mark the base of the older valley.

At this time no extra-local water source has been identified. The most likely areas for such input are along the Battle and Middle Creek Channels. As previously discussed, the Battle Creek channel truncates the channels that cross Merryflat, and connects to a buried valley immediately to the south of the channel. The Middle Creek channel contains slip-off slopes and other evidence of gradual incision and begins in a hanging valley that formed during a later advance. Furthermore, the largest increase in channel cross-sectional area is between Ravenscrag and Eastend and coincides with the junction of the Palisades Coulee. The Palisades Coulee transported meltwater from the south, northwards across slope, and then into the Frenchman channel. There is no indication that a channel from the south or southwest transported water to Palisades Coulee. The Palisades Coulee begins suddenly in the area north of the Old Man On His Back Plateau where there is some evidence for a lake(s) in the area surrounding the Palisades Coulee. This lake may have extended westward to the coalescence zone between the East and West Ice lobes to connect to Lake Cypress. A large glaciolacustrine system may therefore have been present on the plain south of the East Block. No large cross-hill channels equivalent to the Battle Creek channel are present in Reaches 2, 3 or 4.

Implications of the catastrophic release of glacial lakes for the formation of shingled scablands throughout Alberta and Saskatchewan.

This paper presents the hypothesis that the Frenchman channel was cut by the catastrophic release of glacial lakes impounded on or around the Cypress Hills of Alberta and Saskatchewan. Most of the incision is believed to have occurred subaerially but it is possible that some segments of the channel were tunnel valleys cut subglacially. An alternate viewpoint is that a subglacial megaflood originating east of Yellowknife, Northwest Territories (Shaw 1983; Shaw and Kvill 1984; Rains et al. 1993) was responsible for the formation of the Frenchman channel (Shaw, pers. comm. 1993). This megaflood also generated much of the landscape in central Alberta which consists of a broad eroded bedrock surface with little glacigenic or glacial debris on it (Rains et al. 1993). Though lacking in glacigenic debris, the area's numerous drumlins and flutes have been interpreted as subglacial megaflood landforms similar to those described by Shaw (1983, 1989), Shaw and Kvill (1984), and Shaw and Sharp (1987).

The area also contains a large number of spillway channels (Fig. 4.17). Kehew and Lord (1987) and this paper link spillways to the catastrophic release of ice-dammed lakes. The possibility that the glacial debris-free terrain in central Alberta was formed when successive ice-dammed lakes burst and scoured the surface will be examined in this section. The basic premise is that the flutes and drumlins were formed in part by subglacial ice movements perhaps linked to a surging ice lobe. These subglacial landforms were then modified by later local subglacial and subaerial glacial lake outbursts. Each glacial lake outburst scoured part of the terrain, forming a scabland. The overlap of successive outbursts created an apparently continuous surface, called here, a shingled scabland.

The evaluation of the shingled scabland hypothesis begins with an examination of landforms in the Eastend area of Saskatchewan. These landforms can be interpreted as partially resulting from a localized subglacial outburst of a large glacial lake north of the Cypress Hills. In the Shaw and Kvill (1984) and Rains et al. (1993) hypothesis many of the flutings and some of the drumlins observed in Alberta were formed by subglacial megafloods (The path of this megaflood(s) is shown on Fig. 4.17 as dashed lines). (This drumlin genesis contrasts with the view that drumlins form by subglacial accumulation and molding of sediments, see Menzies and Rose, 1989, for a complete discussion of the traditional hypotheses regarding drumlin formation.) In North Dakota elongated drumlins interpreted to have formed as subglacial accumulations behind subglacial obstacles have been described by Bluemle et al. (1993). Dreimanis (1992) described flutes (an elongated form of drumlin?) forming behind obstacles beneath glaciers in Iceland. The flutes had till cores with washed and sorted outer layers of sand and gravel. Crescentic scours were observed around some of the obstacles and flutes. The sand and gravel outer layers and the scour marks were interpreted to have been produced by sub-glacial water flows that eroded and winnowed the surface of the flute. Vreeken (1991) stated that the Swift

Current Plateau had features indicative of subglacial meltwater erosion but did not link these to regional megafloods.

The drumlins in the Dollard drumlin field on the Swift Current plateau consist of sand and gravel layers that mantle ice-thrust bedrock cores. Shaw (pers. comm., 1993).stated that the sand and gravel was deposited from the far-traveled megaflood described in Rains et al. (1993). Crescentic scours and other erosional forms associated with flowing water are also found on or near the drumlins were also incised by the megaflood (Shaw pers. comm., 1993).

An alternative to the non-local megaflood hypothesis is that the Dollard drumlins have a polygenetic origin. Fieldwork north of the Cypress Hills indicates that during the Late Wisconsinan maximum, the ice retreated an unknown distance from its maximum position both north and south of the Cypress Hills (Chapter 6, this volume). The ice then readvanced, thrusting the bedrock and sediment observed in the cores of the Dollard drumlins. Near many of the drumlins in an up-ice direction, are many shallow lakes. These lakes may have been the source areas for the bedrock found in the cores of some of the drumlins. These lakes and drumlin cores would be broadly analogous to the hill-hole pairs described by Bluemle (1970) and Clayton and Moran (1974). These ice-thrust materials would have been the obstacles behind which subglacial debris accumulated, forming the Dollard drumlins. This mechanism is similar to that described by Bluemle et al (1993) and was one of the initial steps in formation of the Dollard drumlin field. During the subsequent wastage of the ice from the area, a large glacial lake formed over the Maple Creek area between the Cypress Hills and the ice to the north (Klassen 1992). It is probable that Eastend Coulee was cut wholly, or in part, by the subglacial drainage of this lake. Initial drainage of this lake subglacially across the Swift Current Creek Plateau and then into the Frenchman channel, would have scoured and winnowed the pre-existing drumlins. Sand and gravel were deposited over the subglacial obstacles and in their lee. The subglacial outburst also eroded the scour marks associated with the drumlins into the plateau surface. The situation is analogous to that described by Dreimanis (1992). The local subglacial drainage of the lake north of the Cypress Hills would therefore produce the same drumlins, coulees, scour-marks, and channel features on the Swift Current plateau as a far-traveled subglacial megaflood.

If the sand and gravel-draped drumlins observed in the Dollard area, and the scours associated with them, were formed by local sub-glacial jokulhaups eroding sediment that had accumulated behind subglacial ice-thrust obstacles, is it possible that these mechanisms were active in other parts of Alberta? Figure 4.16 illustrates the distribution of ice lobes during the Late Wisconsinan as determined by Shetsen (1984).

The Central Lobe of the diagram corresponds to the West Lobe of this paper. It originated in the area of Yellowknife, Northwest Territories (Shetsen 1984) and affected much of central Alberta. The Central Ice Lobe of Shetsen (1984) reached the West Block of the Cypress Hills as indicated by the oolitic ironstone erratics from east of Great Slave Lake present only in areas covered by the West Lobe (Klassen 1991). The West Lobe was very active and may have surged several times (Chapter 6, this volume). The surging may reflect deformation of the soft bedrock beneath the ice (Clayton et al. 1985). The rapid movement of the lobe could have easily thrust bedrock and generated subglacial streamlined features (drumlins and flutes) such as the flutes in the Brooks area of Alberta identified by Campbell (1993) as forming by subglacial meltwater erosion. These flutes are also similar to those interpreted to have formed in the coalescence zone between ice lobes in the Canadian Arctic (John England, personal communication 1993). Because the Brooks flutes are close to the coalescence zone of the West and Central Lobes of Shetsen (1984), they could easily be interpreted as subglacially eroded features and not the products of subglacial megaflood erosion. Once formed (in this case by subglacial erosional processes), the drumlins and flutes in central Alberta could be modified by subglacial drainage events. The question is whether there were several smaller localized outbursts or one or two extra-local megafloods.

The ice of the Late Wisconsin glaciation advanced upslope over much of Alberta. When this ice was melting normal drainage to the northeast was blocked by the continental ice. This impounded numerous glacial lakes in front of the melting ice sheet as it wasted downslope (for example: Stalker 1960, 1973; St. Onge 1972; Shetsen 1984; Proudfoot 1985; Kulig 1985). Spillways leading from many of these glacial lakes have been identified but additional spillways formed by the catastrophic outburst of glacial lakes are continually being identified on the Prairies (see spillway section). The Frenchman, South Saskatchewan, Red Deer, and probably the Battle River and Meeting Creek channels of central Alberta are catastrophically incised spillways. If some of these spillways are linked to subglacial drainage events, as the research by Boothe and Hallet (1992) suggests that they should be, then some of the spillways in central Alberta would be tunnel valleys, scoured and eroded by local subglacial outbursts. Modification of subglacial landforms (ice-thrust bedrock, drumlins, and flutes) by local subglacial outburst(s) in manners similar to those described for the formation of the Dollard drumlin field, could have produced many of the landforms observed in central Alberta. Local subglacial outbursts would produce the large scale fluvial scour marks associated with the landforms.

Bretz (1923, 1981) was the first to describe scablands eroded and scoured by glacial lake outbursts. Bretz's initial observations on the scablands in Washington have been supplemented by observations from many new areas where impounded glacial lakes have catastrophically burst (see spillway section). Subglacial bursting of some of the lakes cut tunnel valleys under the ice and spillways where the outburst became subaerial. The large number of ice-contact lakes and spillways identified in Alberta indicates that much of the terrain of central Alberta could have been affected by these outbursts. Flutes and drumlins close to these spillways would be eroded, have sorted sediments deposited on their outer surfaces, and be linked to fluvial scour marks. The crescentic scours and the sand and gravel deposits and other meltwater erosional features associated with drumlins and flutes (Shaw 1983, 1988; Shaw and Gilbert 1990; Shaw and Kvill 1984) throughout central Alberta would therefore reflect modification of these landforms by localized subglacial jokulhaups. Incorrect interpretation of the subglacial components of these spillways could lead to the interpretation that the widespread distribution of subglacial outburst features reflects one or more extra-local subglacial megaflood events.

The total volume of the outbursts, as discussed before, may not have been confined to the spillway. Extension of the outburst beyond the inner channel of the spillway would have scoured and eroded the area around these channels removing the glacial deposits there. The nature and width of the outer erosional zones associated with these channels are not known at this time. Particularly susceptible are the areas below abrupt turns in the spillways where the outbursts could readily breakout from the channel (Fig. 4.17). As these areas were often unglaciated, the outbursts would have been extremely erosive and scablands with bedrock at the surface would form. If the surging of the ice had already removed much of the basal debris and eroded streamlined landforms into the bedrock then the glacial lake outbursts would not have to remove large quantities of sediment. The exact nature of the surface beneath surging ice lobes is unknown at this time. Research on ice streams in the Antarctic (Alley et al. 1986, 1987a, b, 1989; Blankenship et al. 1987) indicates a zone of deforming sediment but the make up of this sediment is uncertain. On Fig. 4.17 many of the spillway paths coincide at least in part with the megaflood flow path as described in Rains et al. (1993). Some such as the Red Deer spillway in their lower reaches cut across the megaflood flow path. Figure 4.17 is a preliminary diagram of small coulees and spillways, many more have been omitted. Future work will involve mapping many of these smaller channels and comparing them to the megaflood flow path. Once this is done the validity of this argument will be reassessed.

In Alberta the bursting of glacial lakes, both subaerially and probably subglacially, has generated many scablands. The tracts of land in central Alberta that are till and glacigenic sediment deficient but contain numerous drumlins and flutes associated with large scale fluvial scour features, could have been formed by the overlap of several smaller scablands. In the overlap areas integrated pathways could have been eroded which connect one scabland area with the next. Thus, overlap of the scablands from several glacial lake outbursts created an apparently continuous scoured terrain: a shingled scabland. The combination of localized sub-glacial outbursts, intense scouring of ice-free areas by the catastrophic outbursts and glaciofluvial deposition and erosion around pre-existing sub-glacial obstacles, would generate the scoured terrain and landform assemblages described by Rains et al. (1993) without input from extra-local sub-glacial megafloods.

The possible sequence of events to form the shingled scabland would be as follows: The West Lobe (Central Lobe of Shetsen 1984) surges scouring the subsurface to bedrock, incising flutes, drumlins and thrusting sub-glacial obstructions. When the ice begins to melt, ice-dammed lakes are impounded along its margin. These ice-dammed lakes drain catastrophically beneath the ice margin and along ice-walled channels creating subglacial tunnel valleys as well as subaerial spillways. Subglacial outbursts not confined to spillways erode, scour and deposit sorted sediment (sand and gravel) around and near flutes, drumlins and other subglacial obstructions. Outbursts not totally confined to the inner channels of spillways scoured to bedrock broad outer zones (see Kehew and Lord 1986, 1987). Erosion was particularly effective at sharp bends where a spillway changed direction and areas where there is little or very thin laterally confining ice. As the ice continued to waste back, new glacial lakes are impounded at successively lower elevations. The new ice-dammed lakes burst in the same manner as the original and new scabland areas are formed. The overlap of old scabland and new scabland generates a shingled scabland. The process is repeated several times resulting in an extensive semi-continuous shingled scabland that contains subaerial and subglacial spillways and flutes, drumlins and other subglacial obstacles that have been altered by subglacial and subaerial outbursts. These landforms are mantled by sorted sediments and are associated with large scale crescentic scour marks and other erosional features linked to subglacial glaciofluvial outbursts. If each impounded lake burst subglacially or subaerially more than once, a very complicated sequence could be produced. (1992) reported that glacial Lake Missoula burst approximately 40 times.

Conclusions

1) The Frenchman channel is a complex Late Wisconsinan spillway rapidly eroded by the catastrophic release of meltwater from lakes on and around the Cypress Hills. The

initial outburst from Lake Graburn triggered the release of the other lakes on the southern slopes. The outburst was confined along much of its length by ice. The channel indicates that ice was south of the Cypress Hills and Wood Mountain Uplands at the Late Wisconsinan maximum.

2) The cross-sectional area of the channel changes in response to the volume of water released and whether the water was confined between ice or bedrock walls.

3) Ice rafting across a large glacial lake extending from Merryflat to the western end of the West block transported erratics beyond the limits of the Late Wisconsinan ice. These erratics were previously interpreted as the evidence for a pre-Late Wisconsinan ice sheet and were labeled the Elkwater drift (Westgate, 1965a, b, 1968, 1972).

4) The source(s) of the waters that incised the channel require further investigation, as evidence for the lakes is poor and subject to other interpretations. No path for the addition of extra-local water has been found.

5) Eastend Coulee formed after the incision of the Frenchman channel following a subsequent readvance. It drained a large glacial lake situated north of the Cypress Hills. Part of this drainage was probably subglacial.

6) The Adams, Battle, and Middle Creek channels were not part of the initial Frenchman channel but are related to later events within and beyond the study area.

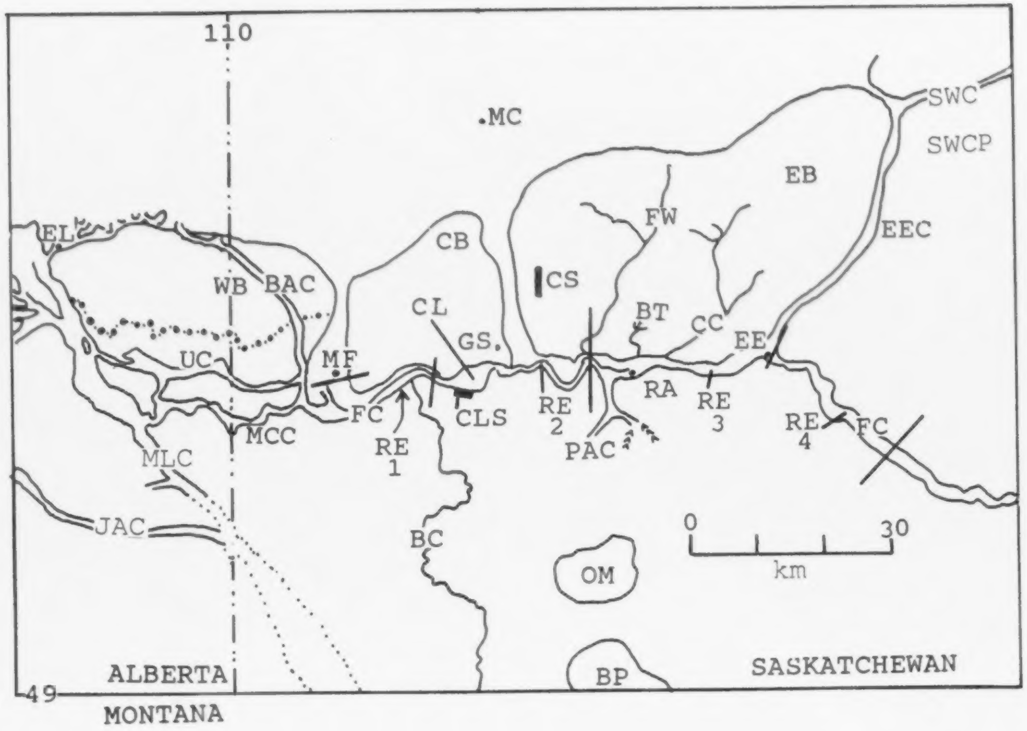
7) The Frenchman channel was used after its initial formation by meltwaters from at least one later glacial advance.

8) Rains et al. (1993) described scoured drift-poor tracts that cover much of central Alberta. This terrain was interpreted as evidence of a subglacial megaflood. An alternative mechanism is that subglacial thrusting of obstacles scoured by repeated subaerial and subglacial bursting of local ice-dammed lakes impounded at different elevations as the Late Wisconsinan Ice retreated formed localized scablands. Coalescence of these local scablands formed an extensive shingled scabland.

Figure 4.1. Map showing the locations, creeks and the boundaries of the four reaches discussed in the text. Letter code: BAC-Battle Creek Channel, BP - Boundary Plateau, BT- Blacktail Creek, CC- Conglomerate Creek, CL - Cypress Lake, CLS - Cypress Lake Sections, CS - Canal Section, EE - Eastend, EEC- Eastend Coulee, EL - Elkwater, FC - Frenchman channel, FW- Fairwell Creek, GS - Gilchrist Section, MF - Merryflat, JAC- Jaydot Coulee, MCC- Middle Creek Channel, MLC- Medicine Lodge Channel, OM - Old Man On His Back Plateau, PAC- Palisades Coulee, RA- Ravenscrag, SWC- Swift Current Coulee, SWCP - Swift Current plateau, UC- Underdahl Channel.

RE1 - reach one from the Merryflat Breach to west Cypress Lake
RE2 - reach two from Cypress Lake to the Palisades Coulee.
RE3 - reach three from Palisades Coulee to Eastend.
RE4 - reach four from Eastend to ~20 km south.

⊏⊐ ⊏⊐ ⊏⊐ ⊏⊐	buried channel
· • · • · • ·	erratics limit on the West Block
>>>>>>	eskers
+——+	reach boundaries
— — — — —	uncertain channel limit.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CLS
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30
km
JAC
BC
OM
ALBERTA
SASKATCHEWAN
49
BP
MONTANA

Figure 4.2. Location of Features and Channels mentioned in the text.
ADC- Adams Creek Channel, BC - Battle Creek, BAC- Battle Creek Channel, BE - Belanger Creek, BP- Boundary Plateau, BT - Blacktail Creek, CC - Conglomerate Creek, CL - Cypress Lake, EE - Eastend, EEC- Eastend Coulee, EK - Elkwater Lake, FRC- Frenchman Channel, FW - Fairwell Creek, JAC- Jaydot Channel, LC - Lyons Creek, MC- Maple Creek, MCC- Middle Creek Channel, MLC- Medicine-Lodge Channel, MF- Merryflat, OM - Old Man On His Back Plateau, PAC- Palisades Coulee, RA - Ravenscrag, RB - Robsart, SU - Sucker Creek, SWC- Swift Current Creek, UC- Underdahl Channel.

Features: F1 - Scours on the interfluve of Adams Creek.
F2 - Breach in the moraine crossing Merryflat.
F3 - The moraine crossing Merryflat.
F4 - Discontinuous ice marginal channels occupied prior to the Frenchman channel's formation.
F5 - boulder-covered scour terrace above the Frenchman channel.
F6 - Streamlined bedforms within the channel.
F7 - Channelettes extending from Palisades Coulee.
F8 - Eskers feeding in the east branch of Palisades Coulee.
F9 - Nested scour channels and gravel pits on the channel's inside bend.

⊏⊐⊏⊐⊏⊐	buried channel
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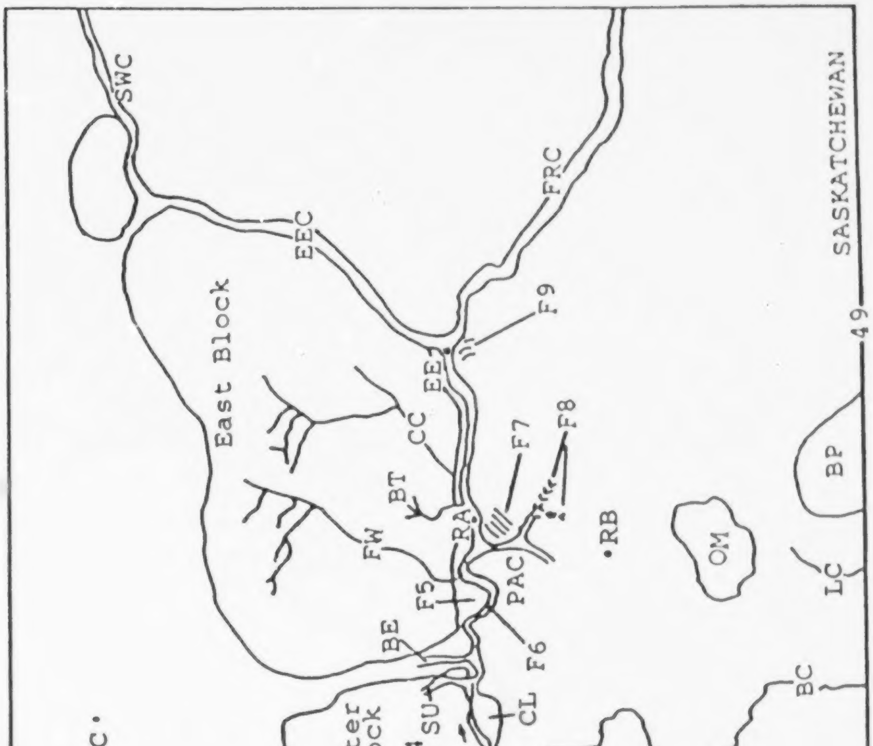
››››››	eskers
◌	moraine crossing Merryflat.


ALBERTA
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MONTANA
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Center Block
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ADC
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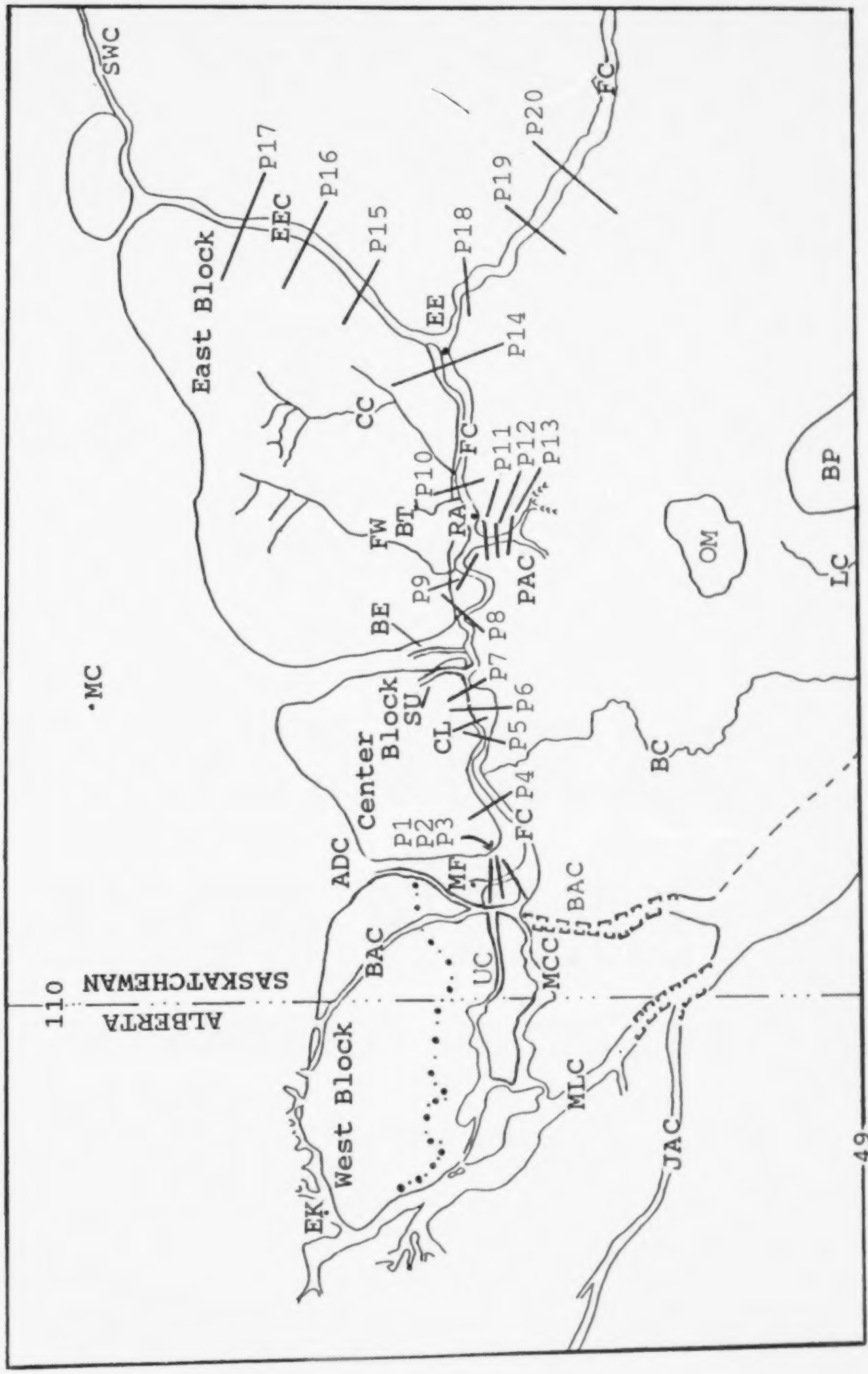
F1
F2
F3
BAC
F4
SU
CL
BC
BE
F6
F5
PAC
LC
OM
RB
FW
RA
BT
BP
F7
F8
CC
EE
F9
EEC
FRC
SWC

Figure 4.3. Map showing the location of profiles discussed in the text. ADC- Adams Creek Channel, BC - Battle Creek, BAC- Battle Creek Channel, BE - Belanger Creek, BP- Boundary Plateau, BT - Blacktail Creek, CC - Conglomerate Creek, CL - Cypress Lake, EE - Eastend, EEC- Eastend Coulee, EK - Elkwater Lake, FC- Frenchman Channel, FW - Fairwell Creek, JAC- Jaydot Channel, LC - Lyons Creek, MC- Maple Creek, MCC- Middle Creek Channel, MLC- Medicine-Lodge Channel, MF- Merryflat, OM - Old Man On His Back Plateau, PAC- Palisades Coulee, RA - Ravenscrag, SU - Sucker Creek, SWC- Swift Current Creek, UC- Underdahl Channel.

⊏⊐⊏⊐⊏⊐	buried channel
· • . • ·	erratics limit on the West Block
>>>>>>	eskers
⊚	moraine crossing Merryflat

P1 - P20 - Profiles used in the text and shown in figures 4.7 and 4.8.

– – – –	uncertain channel margin


SWC
FC
P17
P16
EEC
P15
East Block
P18
EE
P19
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P14
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P10
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ADC
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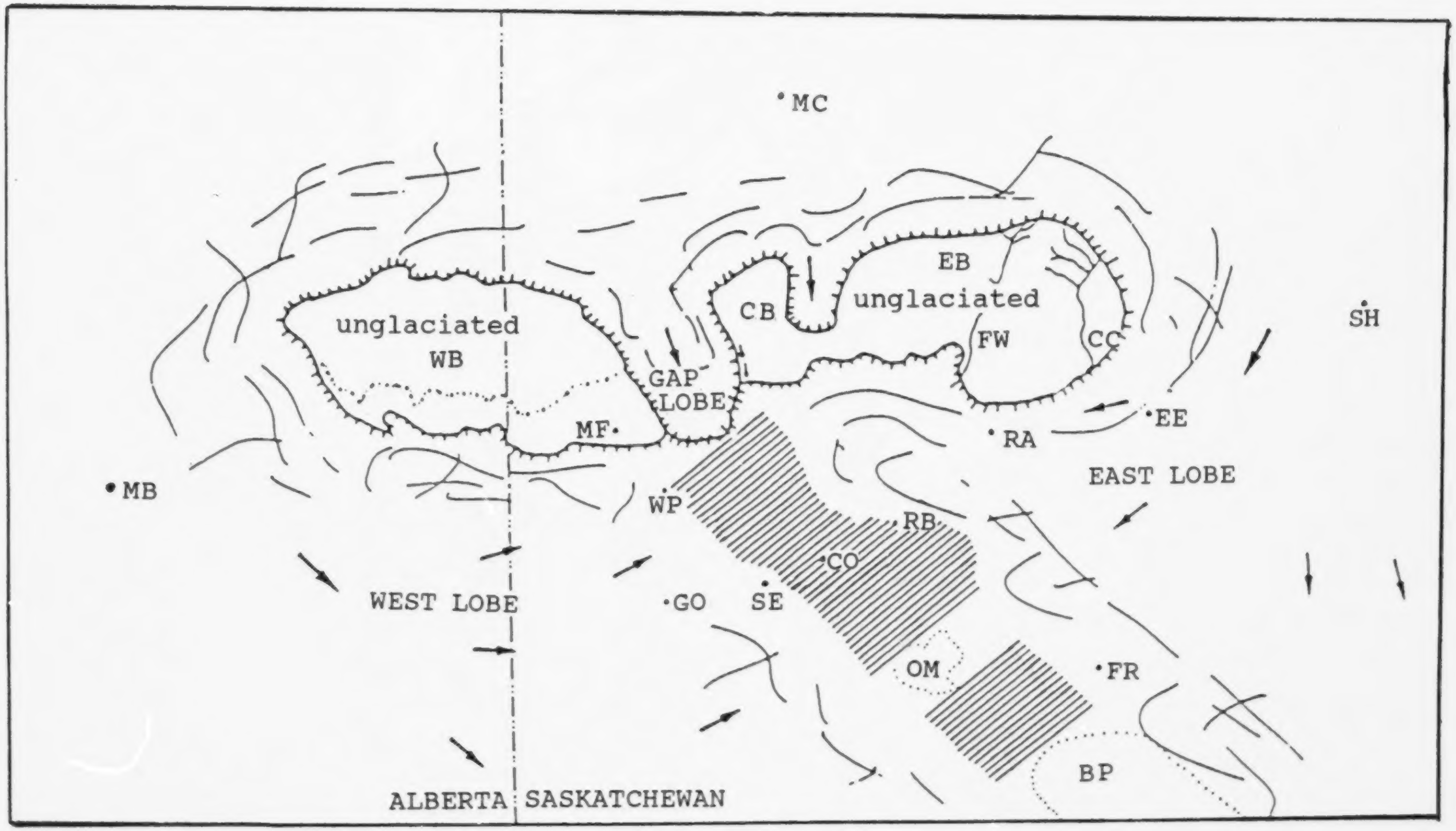
UC
110
SASKATCHEWAN
ALBERTA
MONTANA
West Block
MLC
JAC
49
EK

Figure 4.4. Map showing the distribution of the three ice lobes in the Cypress Hills area at the glacial maximum at ~20, 000 BP. The highest portion of the Old Man On His Back Plateau was only covered by the East Lobe due to the cliff-like nature of its southwestern flank which prevented the West Lobe from ascending onto it. The distribution of the two lobes on the Boundary Plateau is uncertain at this time.

BP- Boundary Plateau, CB- Center Block, CC - Conglomerate Creek, CO- Consul, EB- East Block, EE - Eastend, FR- Frontier, FW - Fairwell Creek, GO- Govenlock, MB- Manyberries, MC- Maple Creek, MF- Merryflat, OM - Old Man On His Back Plateau, RA - Ravenscrag, RB- Robsart, SE- Senate, SH- Shaunavon, WB- West Block, WP- West Plains.

· • · • ·	erratics limit on the West Block
—+—	ice
⋯	ice-covered uplands
➞	ice flow direction
┬┬┬	ice margin
┬ ┬	ice margin uncertain
//////	coalescence zone of the East and West Ice Lobes
→	small discontinuous ice marginal meltwater channels


MC
unglaciated
WB
EB
unglaciated
CB
FW
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SH
GAP
LOBE
MF
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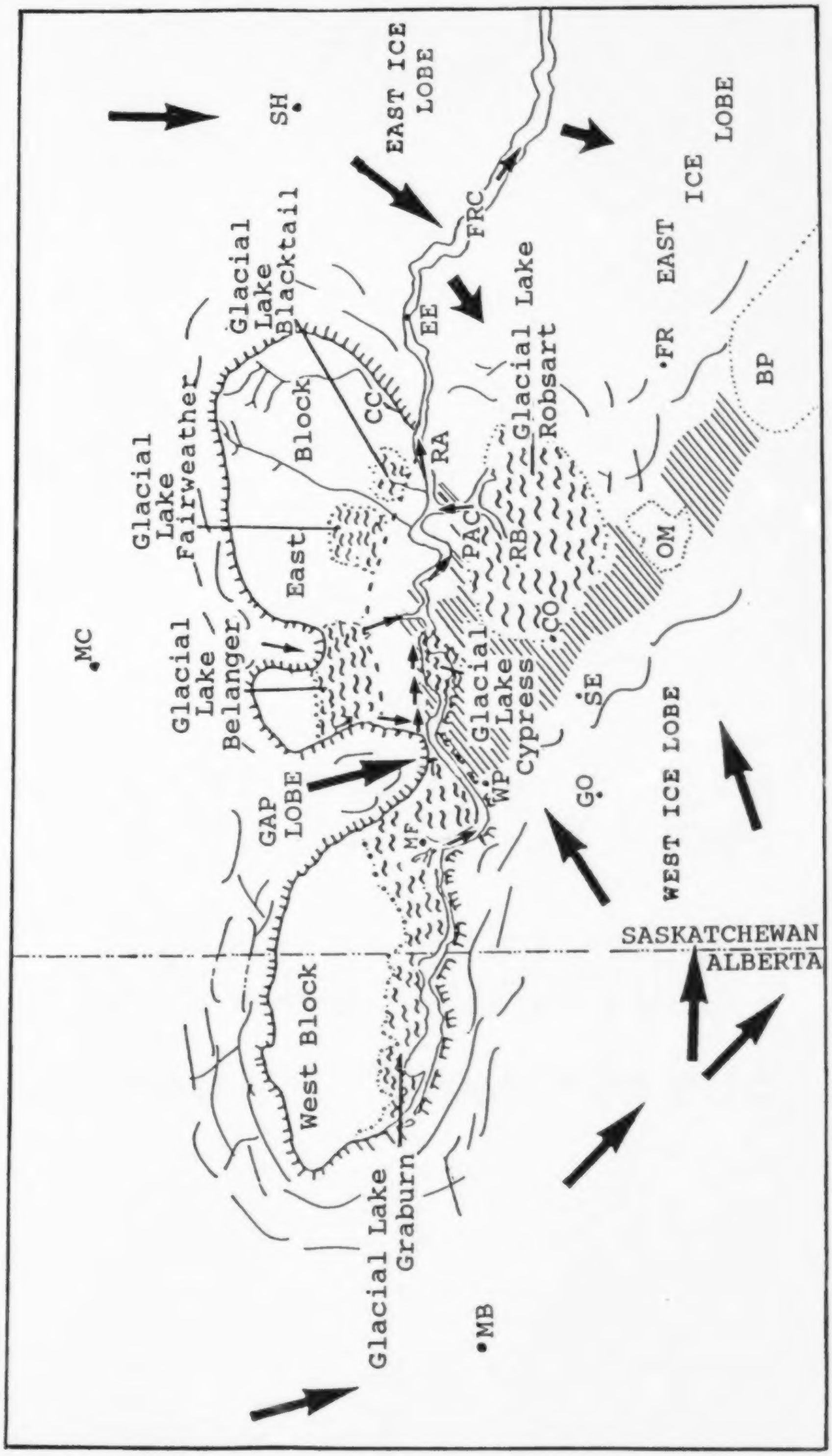
MB
WP
EAST LOBE
RB
CO
WEST LOBE
GO
SE
OM
FR
BP
ALBERTA
SASKATCHEWAN

Figure 4.5. Map showing the possible lakes impounded on or near the south flanks of the Cypress Hills at the beginning of wastage from the Late Wisconsinan Maximum at ~18,000 BP.

BP- Boundary Plateau, CC - Conglomerate Creek, CO- Consul, EE - Eastend, FR- Frontier, FRC - Frenchman Channel, FW - Fairwell Creek, GO- Govenlock, MB- Manyberries, MC- Maple Creek, MF- Merryflat, OM - Old Man On His Back Plateau, RA - Ravenscrag, RB- Robsart, SE- Senate, SH- Shaunavon, WP- West Plains.

·•·•·•·	erratics limit on the West Block
~⊥~⊥~	ice
◌	ice-covered uplands
➡	ice flow direction
┬┬┬┬	ice margin
┬ ┬ ┬	uncertain ice margin
//////	coalescence zone of the East and West Ice Lobes
≈ ≈≈	glacial lakes
⁚≈≈⁚	uncertain glacial lake margins
➔ ➔	water flow direction
→ →	small discontinuous ice marginal meltwater channels


Glacial Lake Graburn
MB
West Block
SASKATCHEWAN
ALBERTA
GAP LOBE
MF
WEST ICE LOBE
GO
WP
Glacial Lake Belanger
Glacial Lake Cypress
SE
MC
CO
East
Block
Glacial Lake Fairweather
PAC
RB
OM
CC
RA
Glacial Lake Robsart
Glacial Lake Blacktail
EE
FR
BP
FRC
EAST ICE LOBE
SH
EAST ICE LOBE

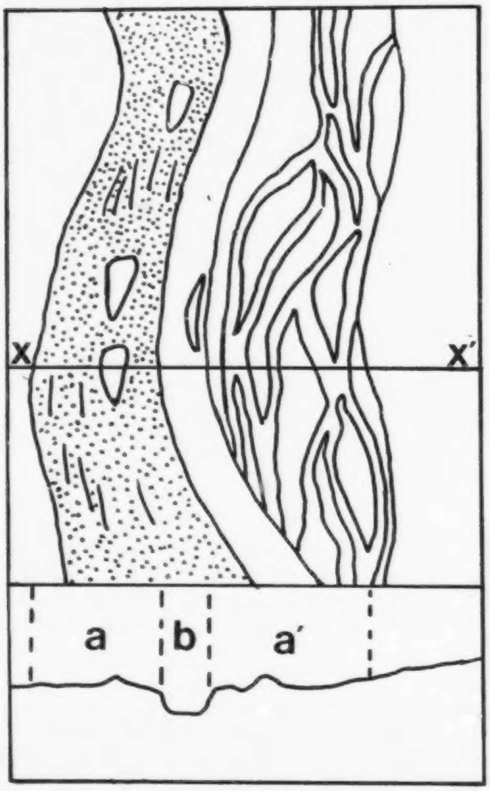
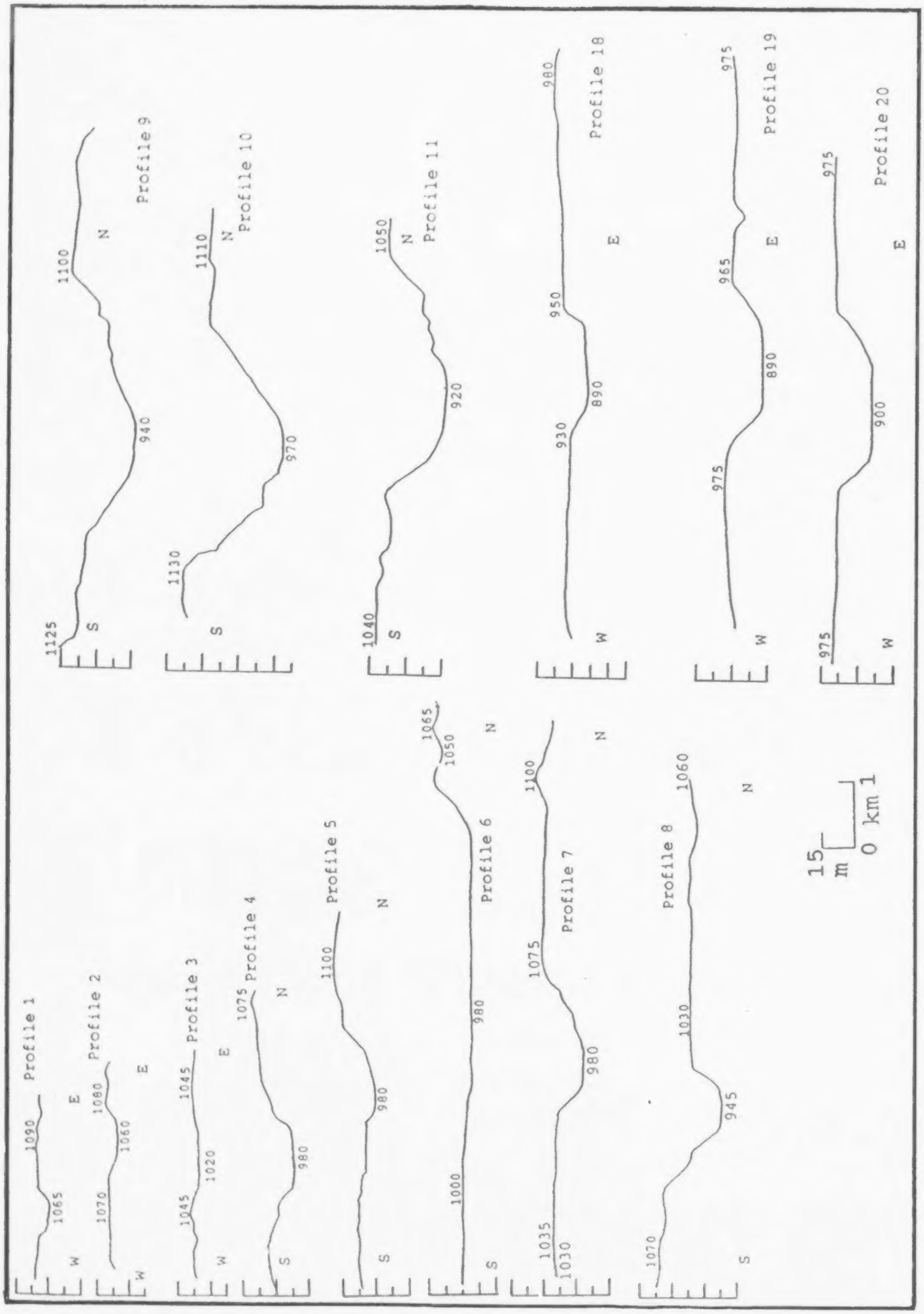


Figure 4.6. Idealized model of the zones and features associated with glacial lake spillways. x - x' cross sectional profile showing the major zones associated with spillways. Zone a, the outer channel contains boulder lags, streamlined residual landforms, and longitudinal grooves. An anastomosing channel system can also be incised in the outer channel before formation of the inner channel (Zone a'). Zone b, the steep-sided inner channel. Deposition is largely confined to large point bars. After Kehew and Lord (1987).

Figure 4.7. Cross-sectional profiles of the Frenchman channel from Merryflat to approximately 20 km south of Eastend, Saskatchewan. Vertical exaggeration 16.7 x.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Profile 7
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Profile 10
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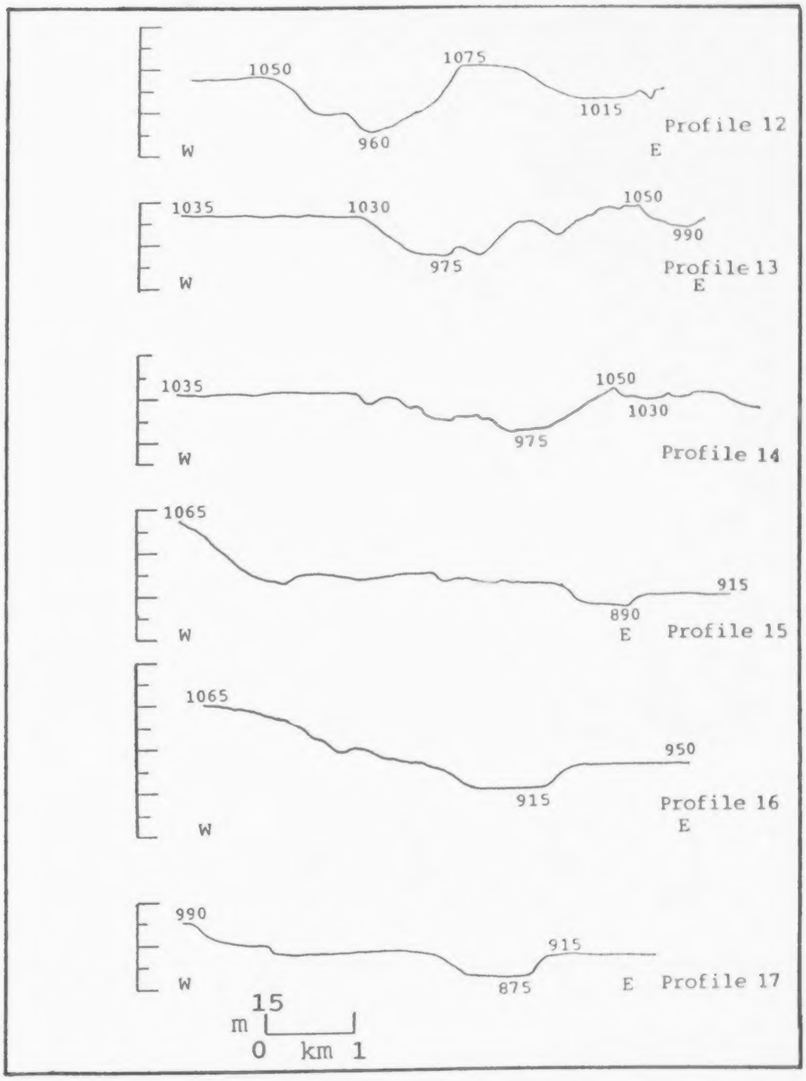


Figure 4.8. Cross-sectional profiles of the Palisades and Eastend Coulees. Vertical exaggeration 16.7 x.

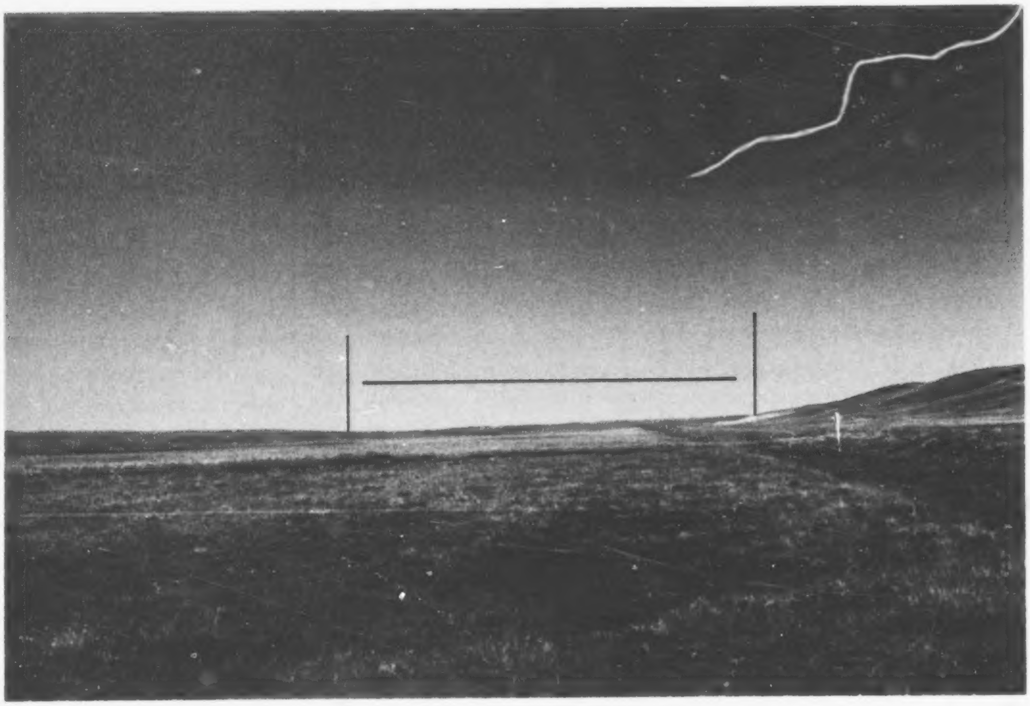


Figure 4.9. Photograph of the breach in the moraine across Merryflat.

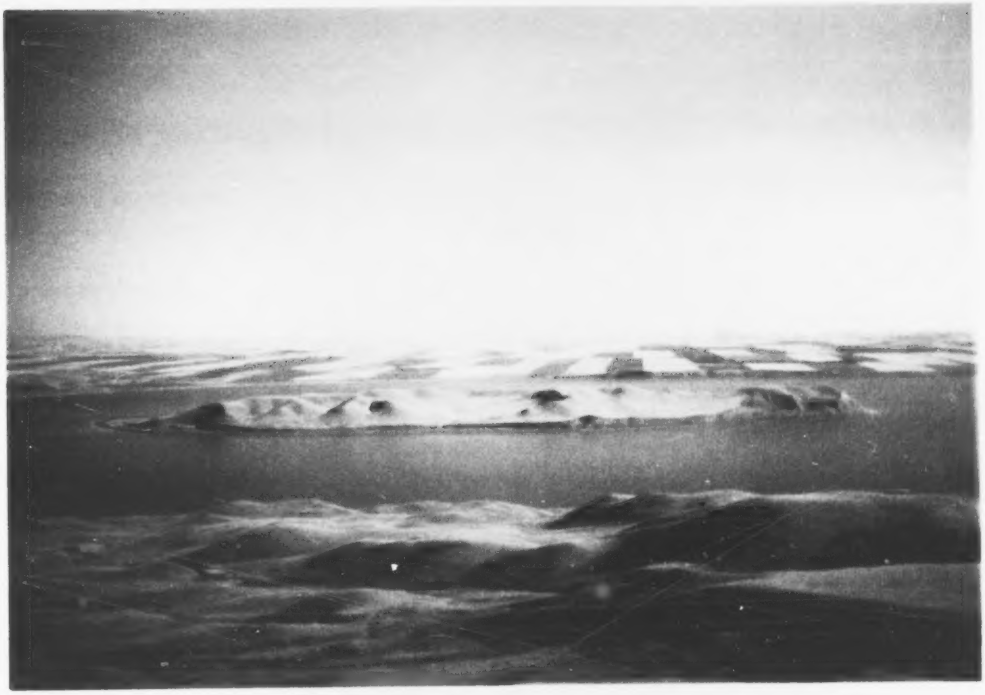


Figure 4.10. The broadening of the channel and the absence of a southern wall at Cypress Lake.

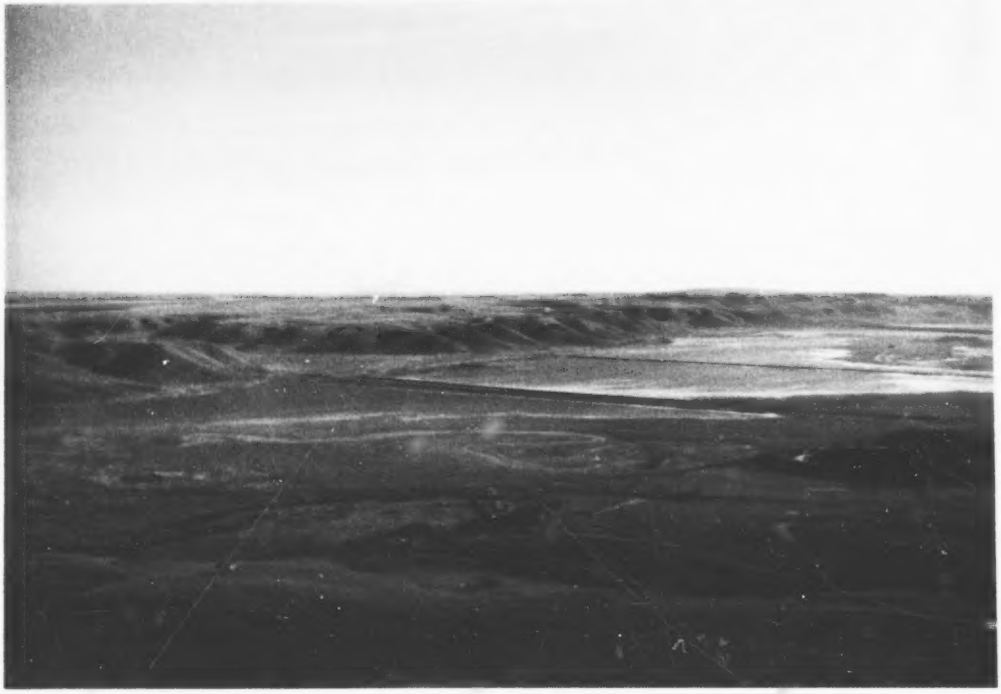


Figure 4.11. The Frenchman channel immediately east of its junction with Sucker Creek.

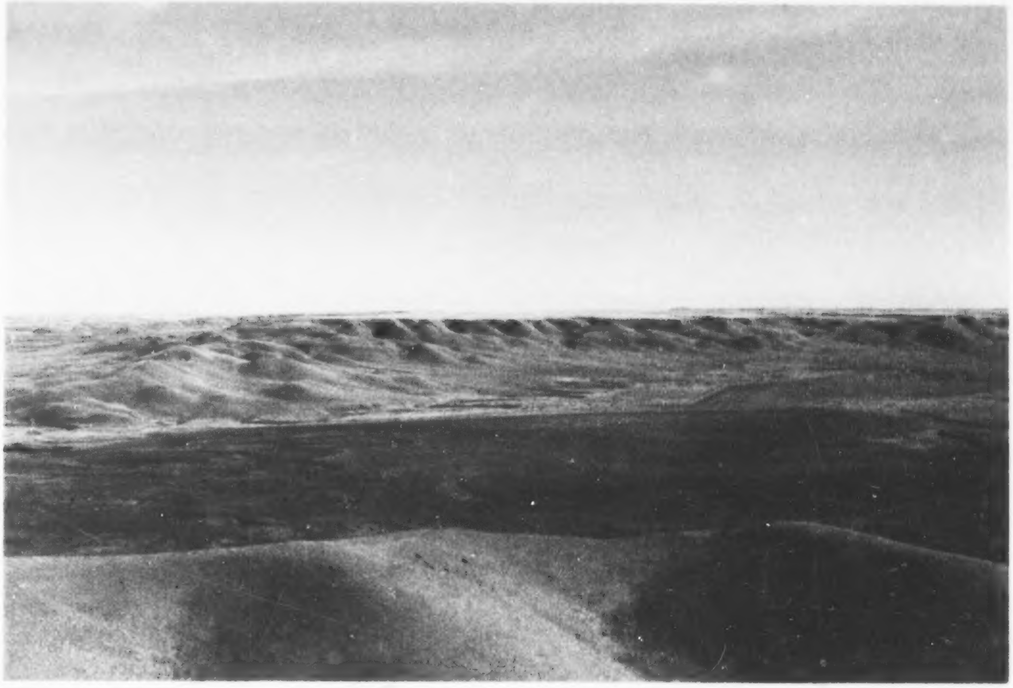


Figure 4.12. The Frenchman channel immediately east of its junction with Belanger Creek.

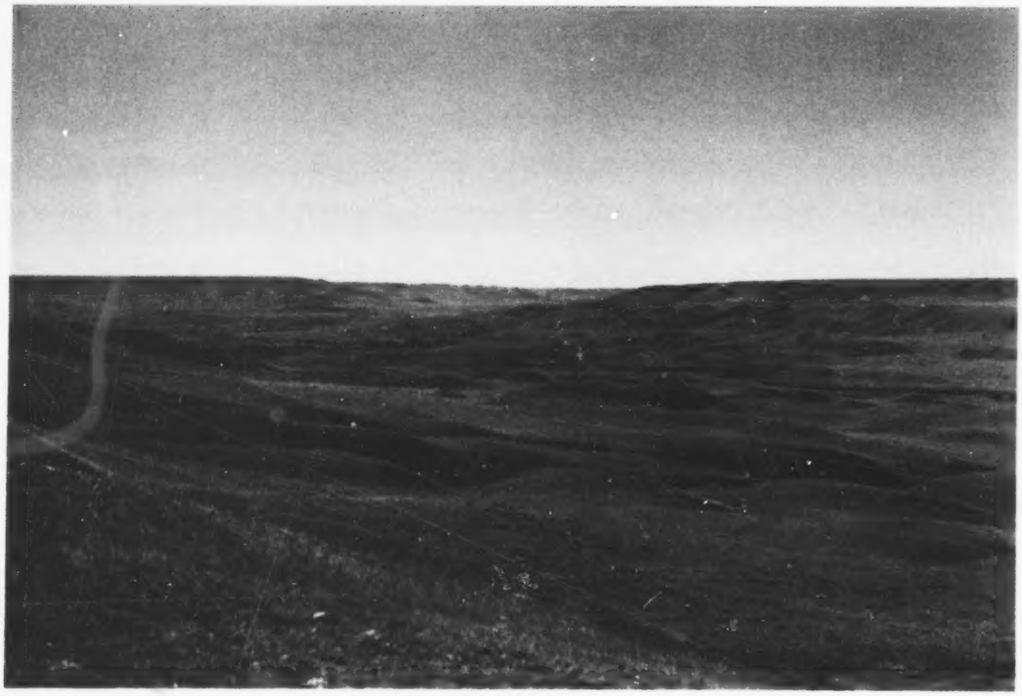


Figure 4.13. Unusual streamlined landforms in the Frenchman channel just before its junction with Palisades Coulee. The landforms may be streamlined residuals formed during the incision of the channel.

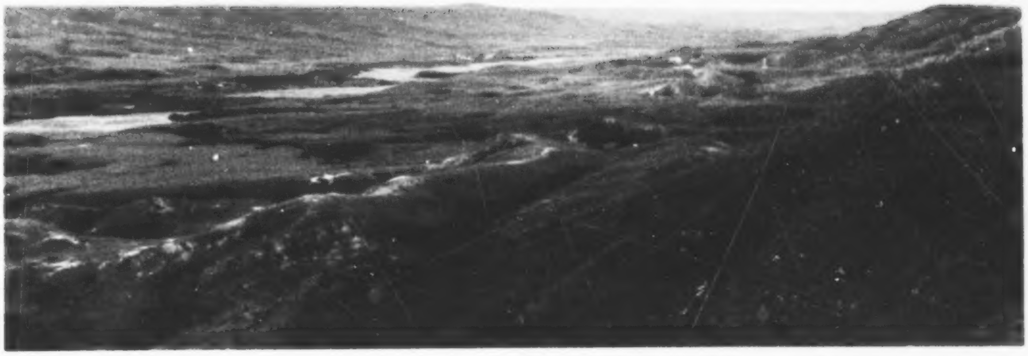


Figure 4.14. View of the Frenchman channel 5 km east of Ravenscrag. Note the slump blocks on the south side of the channel.

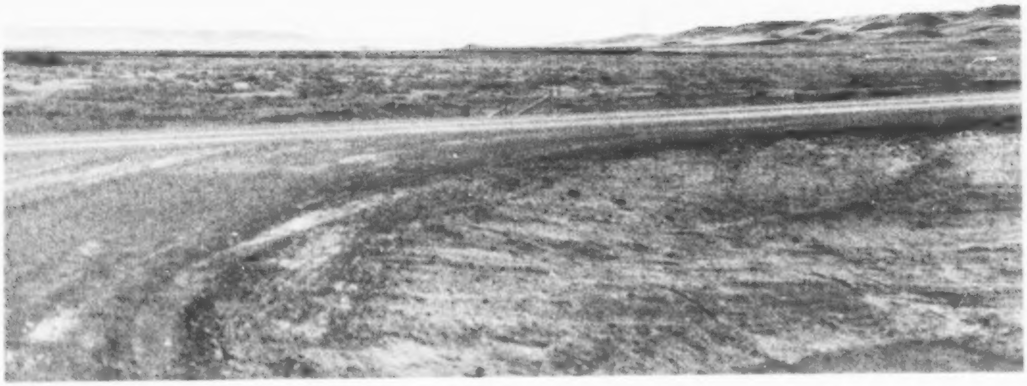


Figure 4.15. The Frenchman channel about 20 km south of Eastend.

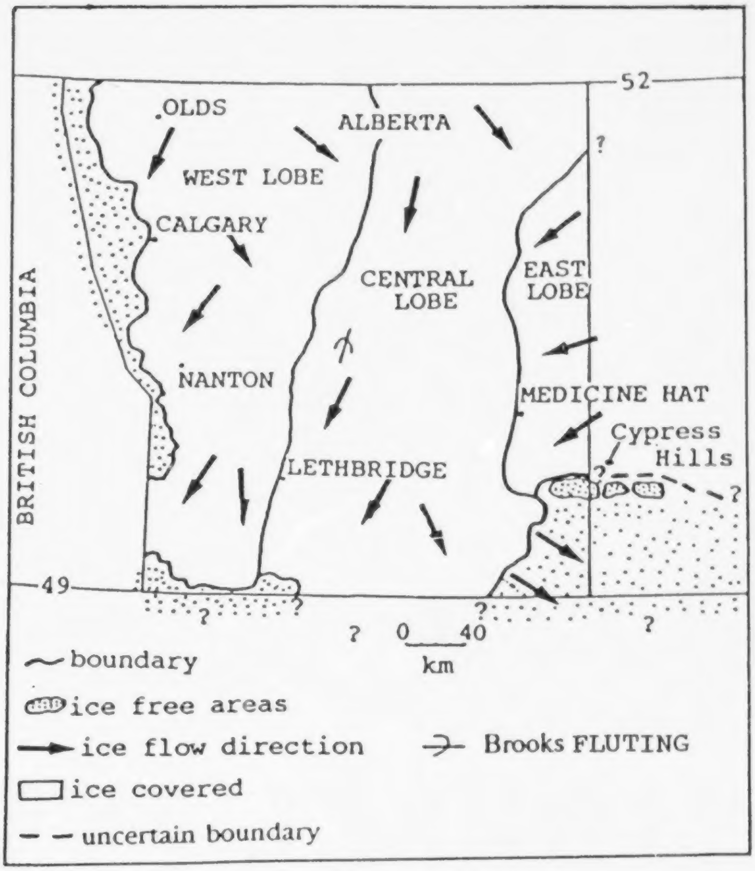
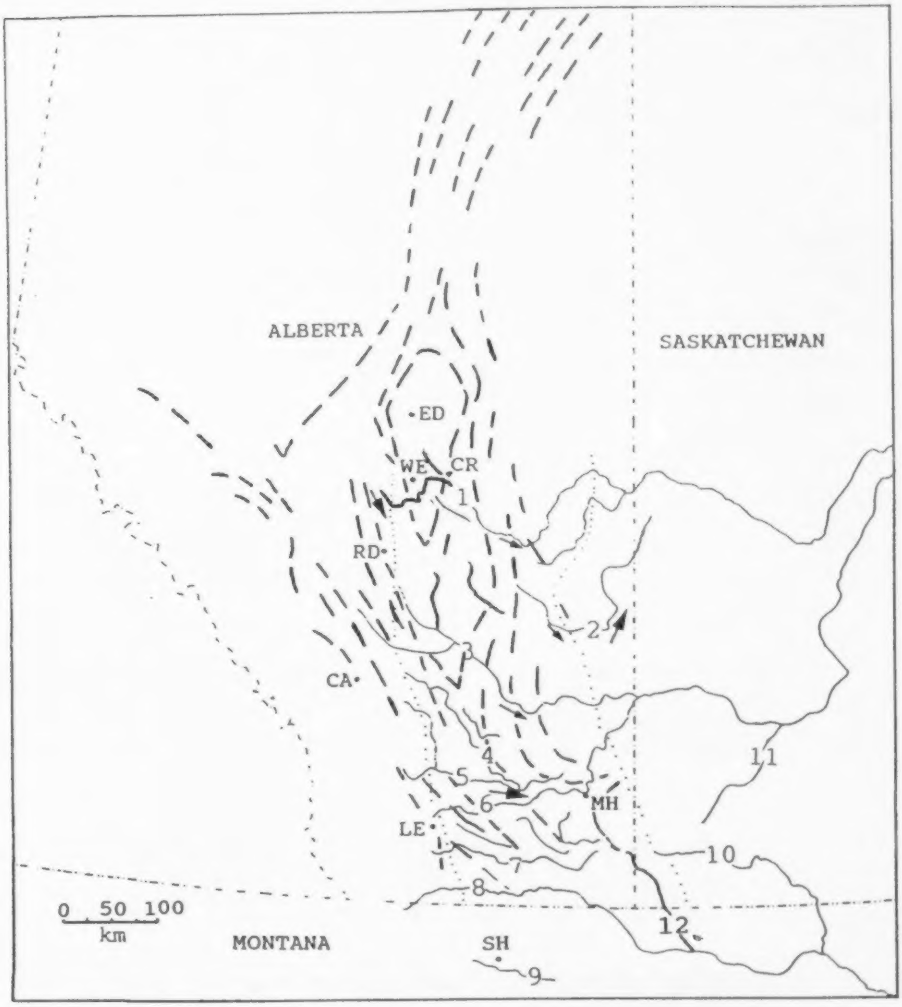

52
49
OLDS
ALBERTA
WEST LOBE
CALGARY
CENTRAL LOBE
EAST LOBE
BRITISH COLUMBIA
NANTON
MEDICINE HAT
Cypress Hills
LETHBRIDGE
?
0 40
km
boundary
ice free areas
ice flow direction
Brooks FLUTING
ice covered
uncertain boundary

Figure 4.16. The distribution of the three ice lobes in Alberta during the Late Wisconsinan after Shetsen (1984). The boundaries of the East Lobe in Alberta and Saskatchewan are uncertain. The valleys and plateaus in the Rocky Mountains of Alberta and British Columbia that were ice free are also unclear.

Figure 4.17. Major spillways in southern Alberta and the possible location scouring by outbursts from impounded glacial lakes. The dotted lines enclose the scabland zone where surficial deposits are restricted and bedrock is at or near the surface. Thin black lines between rivers and creeks are coulees and spillways that are dry. Many more such discontinuous channels are found within the scabland zone. The overlap of outbursts transported along these channels could readily form overlapping scoured scabland surfaces. This forms the shingled scabland referred to in the text. The megaflood flowpath of the megaflood (Rains et al. 1993) is shown by the large dashes. The arrows mark the areas where a sudden course change could allow outbursts to scour areas outside the spillways. CA - Calgary, CR - Camrose, ED - Edmonton, LE - Lethbridge, MH - Medicine Hat, RD - Red Deer, SH - Shelby, WE - Wetaskiwin. 1 - Battle River, 2 - Sounding Creek, 3 - Red Deer River, 4 - Bow River, Little Bow River, 6 - Old man/South Saskatchewan River, 7 - Etzikom Coulee, 8 - Milk River, 9 - Marias River, 10 - Frenchman River, 11- Swift Current Creek, 12 Medicine Lodge Coulee.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References cited.

Aber, J.S., 1991, The glaciation of northeastern Kansas: Boreas, v. 20, p. 297-314.

Alley, R.B., Blankenship, D.D., Bentley, C.R., and Rooney, S.T., 1986, Deformation of till beneath ice stream B, West Antarctica: Nature, v. 322, p. 57-59.

Alley, R.B., Blankenship, D.D., Bentley, C.R., and Rooney, S.T., 1987a, Till beneath ice stream B, 3, Till deformation: evidence and implications: Journal of Geophysical Research, v. 92, p. 8921-8930.

Alley, R.B., Blankenship, D.D., Rooney, S.T., and Bentley, C.R., 1987b, Till beneath ice stream B, 4, a coupled ice-till flow model: Journal of Geophysical Research, v. 92, p. 8931-8940.

Alley, R.B., Blankenship, D.D., Rooney, S.T., and Bentley, C.R., 1989, Sedimentation beneath ice shelves - the view from ice stream B: Marine Geology, v. 85, p. 101-120.

Ashley, G.M., 1989, Classification of glaciolacustrine sediments, *in* Goldthwait, R.P., and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam, p. 243-260.

Baker, V. R., Benito, G., and Rudoy, A.N., 1993, Paleohydrology of Late Pleistocene superflooding, Altay Mountains, Siberia: Science, v. 259, p. 348-350.

Blankenship. D.D., Bentley, C.R., Rooney, S.T., and Alley, R.B., 1986., Seismic measurements reveal a porous layer beneath an Antarctic ice stream: Nature, v. 322, p. 54-57.

Bluemle, J.P., 1970, Anomalous hills and associated depressions in central North Dakota, Geological Society of America, Abstracts with Program, v. 2, p. 325-326.

Bluemle, J.P. , Lord, M.L., and Hunke, N.T., 1993, Exceptionally long narrow drumlins formed in subglacial cavities, North Dakota: Boreas, v. 22, p. 15-24.

Booth, D.B., and Hallet, B., 1993, Channel networks carved by subglacial meltwater: observations and reconstruction in the eastern Puget Lowland of Washington: Geological Society of America Bulletin, v. 105, p. 671-683.

Brennand, T.A., and Shaw, J., 1994: Tunnel channels and associated landforms, south-central Ontario: their implications for ice sheet hydrology, Canadian Journal of Earth Sciences, v. 31, 505-522.

Bretz, J.H., 1923, The Channeled Scablands of the Columbia Plateau: Journal of Geology, v. 31, p. 617-649.

Bretz, J.H. 1981, Part 1: an outrageous hypothesis, *in* Baker, V. R., ed, Catastrophic Flooding, The Origin of the Channeled Scablands, Dowden, Hutchingson and Ross, Stroudsburg, Penn. p. 18-58.

Christiansen, E.A. 1979, The Wisconsinan glaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Christiansen, E.A., and Sauer, E.K. 1988, Age of the Frenchman Valley and associated drift south of the Cypress Hills, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 25, p. 1703-1708.

Clayton, Lee, 1983, Chronology of Lake Agassiz drainage to Lake Superior, *in*: Teller, J.T. and Clayton, Lee, eds., Glacial Lake Agassiz, Geological Association of Canada Special Paper, v. 26, p. 290-307.

Clayton, Lee, and Cherry, J.A., 1967, Pleistocene supraglacial and ice-walled lakes of west-central North America, in Clayton, Lee and Freers, T. F., eds., Glacial Geology of the Missouri Coteau and adjacent areas, North Dakota Geological Survey Miscellaneous Series 30, p. 47-52.

Clayton, Lee., and Moran, S., 1974, A glacial-process form model, *in* Coates, D.R., (ed.), Glacial Geomorphology, Binghampton, New York, SUNY-Binghampton Publications in Geomorphology, p. 89-119.

Clayton, Lee and Moran, S.R., 1982, Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55-82.

Clayton, Lee, Teller, J.T., and Attig, J.W., 1985, Surging of the southwestern part of the Laurentide Ice Sheet: Boreas, v. 14, p. 235-241.

Dreimanis, A., 1992, Water-eroded crescentic scours and furrows associated with subglacial flutes at Breidamerkuljokull, Iceland: Boreas, v. 22, p. 110-112.

Dyke, A.S., and Prest, V. K., 1987a, Late Wisconsinan and Holocene history of the Laurentide ice sheet: Geographie Physique et Quaternaire: v. 41, p. 237-263.

Dyke, A.S., and Prest, V. K., 1987b, The Late Wisconsinan Maximum, Geological Survey of Canada map 1702a, scale 1:5,000,000.

Elfstrom, A., 1987, Large boulder deposits and catastrophic floods: Geografiska Annaler, v. 69A, p. 101-121.

Embleton, C. and King, C.A.M., 1975, Glacial Geomorphology, 2nd ed. London, Edward Arnold. 573 pp.

Evans, D.J.A., and Campbell, I.A., 1992, Glacial and postglacial stratigraphy of Dinosaur Provincial Park and surrounding plains, southern Alberta, Canada: Quaternary Science Reviews, v. 11, p. 535-555.

Glen, J.W., 1954, The stability of ice dammed lakes and other water-filled holes in glaciers: Journal of Glaciology, v. 2, p. 316-318.

Howarth, P. J., 1968, A supra-glacial extension of an ice dammed lake, Tunsbergdalsbreen, Norway: Journal of Glaciology, v. 51, p. 413-419.

Jarret, R.D., and Malde, H.E., 1987, Paleodischarge of the Late Pleistocene Bonneville Flood, Snake River, Idaho, computed from new evidence; Geological Society of America Bulletin, v. 99, p. 127-134.

Kehew, A.E., 1982, Catastrophic flood hypothesis for the origin of the Souris spillway, Saskatchewan and North Dakota: Geological Society of America Bulletin, v. 93, p. 1053-1058.

Kehew, A.E., 1993, Glacial lake outburst erosion of the Grand Valley, Michigan, and impacts on glacial lakes in the Lake Michigan Basin: Quaternary Research, v. 39. p. 36-44.

Kehew, A.E., and Clayton, L., 1983, Late Wisconsinan floods and the development of the Souris-Pembina spillway system in Saskatchewan, North Dakota, and Manitoba, *in* Teller, J.T. and Clayton, Lee, eds., Glacial Lake Agassiz, Geological Association of Canada Special Paper 26, p. 231-244.

Kehew, A.E., and Lord, M.L., 1986, Origin and large-scale erosional features of glacial-lake spillways in the Northern Great Plains: Geological Society of America Bulletin, v. 97, p. 162-177.

Kehew, A.E., and Lord, M.L., 1987, Glacial-lake outbursts along the mid-continent margins of the Laurentide ice-sheet, *in* Mayer, L., and Nash, D., eds., Catastrophic Flooding, Allen and Unwin, Boston, p. 95-120.

Klassen, R.W. 1984, Quaternary geology of southwestern Saskatchewan., *in* Current Research, Part A, Geological Survey of Canada. Paper 84-1A, p. 641-642.

Klassen, R.W., 1991, Surficial geology and drift thickness, Cypress Lake (72F) Saskatchewan. Geological Survey of Canada, Map 1766A (1:250,000).

Klassen, R.W. 1992, Nature, origin and age relationships of landscape complexes in southern Saskatchewan: Geographie physique et Quaternaire, v. 46, p. 361-388.

Klassen, R.W., and Vreeken, W.J., 1985, Quaternary Geology of southwestern Saskatchewan, *in* Current Research, Part A, Geological Survey of Canada, Paper 85-1A, p. 187-192.

Klassen, R.W. and Vreeken, W.J. 1987, The nature and implications of surface tills and post-till sediments in the Cypress Lake area, Saskatchewan, *in* Current Research, Part A, Geological Survey of Canada, Paper 87-A, p. 111-125.

Kor, P. S.G., Shaw, J., and Sharpe, D.R., 1991, Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario: a regional review: Canadian Journal of Earth Sciences, v. 28, p. 623-642.

Kulig, J.J., 1985, A sedimentation model for the deposition of glacigenic deposits in central Alberta. [M.S. thesis], Edmonton, Alberta, University of Alberta, 245 p.

Lindsay, J.F., 1966, Observations on the level of a self-draining lake on the Casement Glacier, Alaska: Journal of Glaciology, v. 45, p. 443-445.

Liverman, D.G.E., 1981, Sedimentology and drainage history of a glacier dammed lake, St. Elias Mountains, Yukon Territory. [M.S. thesis], Edmonton Alberta, University of Alberta, 221p.

Lord, M.L., and Kehew, A.E., 1987, Sedimentology and paleohydrology of glacial-lake outburst deposits in southeastern Saskatchewan and northwestern North Dakota: Geological Society of America Bulletin, v. 99, p. 663-673.

Maizels, J., 1989, Sedimentology, paleoflow dynamics and flood history of Jokulhlaup deposits: paleohydrology of Holocene sediment sequences in southern Iceland sandur deposits.

Mathews, W.H., 1965, Two self-dumping ice-dammed lakes in British Columbia: Geographical Review, v. 55, p. 46-52.

Matsch, C.L., 1983, River Warren, the southern outlet to glacial Lake Agassiz, *in* Teller, J.T. and Clayton, Lee, eds., Glacial Lake Agassiz, Geological Association of Canada Special Paper 26, p. 231-244.

Menzies, J., and Rose, J., 1989, Subglacial bedforms - drumlins, Rogen Moraine and associated bedforms: Sedimentary Geology, v. 62, 117-430.

Pardee, J.T., 1942, Unusual currents in Glacial Lake Missoula, Montana: Bulletin of the Geological Society of America, v. 53, p. 1569-1600.

Proudfoot, D.N., 1985, A lithostratigraphic and genetic study of Quaternary sediments in the vicinity of Medicine Hat, Alberta. [Ph.D. thesis], Edmonton, Alberta, University of Alberta, 345 p.

Rains, B., Shaw, J., Skoye, R., Sjogren, D., and Kvill, D., 1993, Late Wisconsinan mega-flood paths in Alberta: Geology, v. 21, p. 323-326.

Rabot, C., 1905, Glacial reservoirs and their outbursts: The Geographical Journal, v. 55, p. 534-548.

St.Onge, D.A., 1972, Sequence of glacial lakes in north-central Alberta: Geological Survey of Canada Bulletin, 213, 16pp.

Shaw, J., 1983, Drumlin formation related to inverted meltwater erosional marks: Journal of Glaciology, v. 29, p. 461-479.

Shaw, J., 1989, Drumlins subglacial meltwater floods, and ocean responses: Geology, v. 17, p. 853-856.

Shaw, J., and Kvill, D., 1984, A glaciofluvial origin for drumlins of the Livingstone Lake area, northern Saskatchewan: Canadian Journal of Earth Sciences, v. 21, p. 1442-1459.

Shaw, J., Kvill, D., and Rains, B., 1989, Drumlins and catastrophic subglacial floods: Sedimentary Geology, v. 62, p. 177-202.

Shaw, J., and Sharpe, D.R., 1987, Drumlin formation by subglacial meltwater erosion: Canadian Journal of Earth Sciences, v. 24, p. 2316-2322.

Shetsen, I. 1984, Application of till pebble lithology to the differentiation of glacial lobes in southern Alberta: Canadian Journal of Earth Sciences, v. 21, p. 920-933.

Stalker, A. MacS., 1960, Surficial geology of the Red Deer-Stettler map area: Geological Survey of Canada Memoir 306, 140p.

Stalker, A. MacS., 1973, Surficial geology of the Drumheller area, Alberta: Geological Survey of Canada Memoir 370, 122 p.

Stone, K.H., 1963, Alaskan ice-dammed lakes: Annals of the Association of American Geographers, v. 53, p. 332-349.

Teller, J.T., and Thorliefson, L.H., 1983, The Lake Agassiz-Lake Superior connection: in Teller, J.T. and Clayton, Lee, eds. Glacial Lake Agassiz. Geological Association of Canada Special Paper 26, p. 261-290.

Thorarinsson, S., 1939, The ice-dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations: Geographisca Annaler, v. 21, p. 216-242.

Vreeken, W.J. 1986, Quaternary events in the Elkwater area of southeastern Alberta: Canadian Journal of Earth Sciences, v. 23, p. 2024-2038.

Vreeken, W.J. 1991, The southern Swift Current Plateau (Saskatchewan): A sub-glacial meltwater erosion surface: Geological Association of Canada Annual Meeting, Toronto, Program with Abstracts, v. 16, p. A129.

Westgate, J.A., 1965a, The surficial geology of the Cypress Hills area, Alberta: Alberta Research Council Preliminary Report 65-2, 6pp.

Westgate, J.A., 1965b, The Pleistocene stratigraphy of the Foremost-Cypress Hills area Alberta: Alberta Society of Petroleum Geologists, 15th Annual Field Conference, Guidebook, Part 1, Cypress Hills Plateau, p. 85-111.

Westgate, J.A., 1968, Surficial geology of the Foremost - Cypress Hills area, Alberta: Research Council of Alberta Bulletin 22, 122 pp.

Westgate, J.A., 1972, The Cypress Hills. *in* Quaternary geology and Geomorphology between Winnipeg and the Rocky Mountains. International Geological Congress, 24th Session Montreal Quebec, Guidebook for Excursion C-22 p. 50-62.

Chapter 5: The role of micromorphology in genetic studies of glacigenic diamicton.

A version of this paper will be submitted for publication in Journal of Quaternary Research.

Introduction

During investigations into the genesis of glacigenic sediments in Alberta and southwest Saskatchewan block samples were collected for the production of thin-sections (Fig. 5.1). A total of 96 longitudinal thin-sections, from three glacial sediment types, subglacial meltout till, subaerial glacigenic debris-flows, and glaciolacustrine diamicton complexes were made. These thin-sections were examined for features characteristic of each depositional environment. The thin-sections enabled close examination of sedimentary structures such as sub-horizontal sand partings, internal banding, and diamicton intraclasts (small generally coarse sand to pebble-sized clasts of clay-rich diamicton lacking phenoclasts), that are not readily observed in field exposures. Recognition of diamicton intraclasts is often critical to understanding the genesis of a glacigenic deposit and making more precise and in-depth genetic interpretations. The objectives of this paper are to document features visible in thin-section that are diagnostic of the three glacial sediment types, to provide information on the sedimentary features most useful in thin-section analysis, and to evaluate the usefulness of micromorphology as a major investigative tool in till genesis studies. Since the terminology for describing thin-sections of unconsolidated material is not often used in geological investigations, definitions of the terms are given in Appendix 1.

Previous work

Thin-sections of till and related deposits have been used for two inter-related purposes: (1) to identify micromorphological features useful for the determination of till genesis and (2) to determine ice-flow direction from examination of the microfabric of the glacial deposits. Sitler and Chapman (1955), the first to use till thin-sections, identified vein structures and microfoliation which they related to shearing during deposition. Korina and Faustova (1964) described reticulate fibrous textures and the concentric orientation of clay around sand grains. They concluded that these features were inherited from the ice structure and preserved after meltout. Van der Meer et al. (1983) described mud and diamicton intraclasts that were deformed when the original deposits were over-ridden by ice. They also described skelsepic plasma fabric in the till matrix and unistrial fabric in shear zones. Kulig (1985) used thin-sections as an aid in determining the origin of sand and silt laminae intercalated with diamicton beds in glaciolacustrine diamicton units located along the Battle River, in central Alberta. Fluid-escape structures, diamicton intraclasts, rhythmite intraclasts, and laminae-phenoclast interrelationships were also described (Kulig 1985). Van der Meer (1987) described mud pebbles, bedding and banding, fluid-escape structures, and several examples of plasma fabrics within till and suggested that the system of thin-section description widely used by soil scientists (Brewer 1976; Brewer and Sleeman 1988) be used for till thin-section analysis. This paper follows this suggestion. Van der Meer and Laban (1990) concluded that the North Sea material they examined was glacial in origin because it possessed very low porosity, contained erratics, and possessed moderate to well-developed skelsepic plasma fabrics. Van der Meer et al. (1992) detailed the micromorphology of Argentine glaciolacustrine sediments and concluded that the sediment lacked a clear "glacial imprint". The glacial imprint was described as high density, few and discontinuous pores, and skelsepic plasma fabrics. The presence of unistrial plasma fabrics, kink bands, and shear zones in thin-sections from drumlins in central Canada led Menzies and Maltman (1992) to conclude the drumlins formed by active subglacial sediment deformation.

Determination of ice-flow direction through microfabric analysis was first undertaken by Ostry and Deane (1963). They concluded that the microfabric of a till closely matched its phenoclast fabric. Subsequent studies (Evenson 1970, 1971; Johnson 1983) confirmed this correspondence. Sitler (1968) concluded that the microfoliation visible in some till thin-sections also paralleled the ice-flow direction.

Methods

Diamicton block samples, of each sedimentary unit as well as blocks containing specific features such as laminae or bedrock stringers, were obtained from the glacigenic sediment undergoing investigation. The oriented block samples were broken along natural planes of weakness. This resulted in samples in which horizontal sedimentary features are off angle. This sampling method was used since stones in the diamicton prevented the cutting square or rectangular blocks from most of the sections. The block samples were oriented by marking them directional arrows for up and letters for front and rear faces, using white paint. Visible structures were also highlighted. The blocks were air dried for several months and then oven-dried at about 50° C for 1 day to minimize features created by clay shrinkage. The dried blocks were impregnated under vacuum with 3M Scotchcast epoxy casting resin then cured overnight in an oven set at 60° C. The impregnated samples were trimmed into 7 by 5 by 1 cm blocks and one side was polished using a Logitech grinder and 9 µm grit. The polished side was then mounted on a frosted glass slide using Epotek mounting medium. An offcut saw removed the bulk of the block and a surface grinder ground the slide to between 75 to 100 µm. A two step finishing process using the Logitech, began with 14.5 µm carbide grit and then 9.5 µm grit to produce a finished thin-section of between 25 to 30 µm thickness.

The blocks and subsequent thin-sections were coded to reduce operator bias during examination. The thin-sections were examined at several magnifications but comparisons between thin-sections were made at the same magnification.

Locations and geologic contexts

Blocks for thin-section production were obtained from three field areas in Alberta and southwest Saskatchewan (Fig. 5.1) where the genesis of the diamictons had been investigated. At the Battle and Lyons creek sections in southwest Saskatchewan, two units, interpreted as a subglacial meltout till unit and an overlying subaerial glacigenic debris-flow complex, were sampled. The UTM grid coordinates for Battle Creek 1 (BC1) are 72F/3,12UXK153287, and for BC2 72F/3,12UXK147379 (LSD: W of 3, T1 R26 S4 1/4 sec. south half of 3 to SW quad. of 6; W of 3, T1 R26 S4 1/4 sec. SW quadrant of 1). The grid co-ordinates of the Lyons Creek sections are 72F/3,12UXK282337 (LC1), 72F/3,12UXK282338 (LC2), 72F/3,12UXK278337 (LC3), 72F/3,12UXK282342 (LC4) (LSD: LC1: West of 3, T1, R25, S14, 1/4 sect., NW quadrant of 14, LC2: T1, R25, S23, 1/4 sect., SW quadrant of 3, LC3: T1, R25, S23, 1/4 sect., NE quadrant of 4, LC4: T1, R25, S23, 1/4 sect., SE and NE quadrants of 11.). The block samples from hummocks in the Cooking Lake moraine (location 83H12UUQ642.1342.2, LSD: T53 R21 S4 1/4 sec. SW quadrant of 4) in central Alberta are from sediment interpreted to be subaerial glacigenic debris-flow complexes (Kulig 1989). The thin-sections from the Cypress Lake, Gilchrist and Canal sections and those exposed along Battle River in central Alberta (Battle River section, 83A/14,12UUP530.0653.3; LSD: W of 4, T45 R22

S31 1/4 sec NE quadrant of 3) were obtained from deposits interpreted to be glaciolacustrine debris-flow complexes, deposited in ice-contact lakes (Kulig 1985; chapter 3, this volume). The UTM grid co-ordinates for the Canal section are 72F/11,12UXK178918 for the south end of the section and 177929 for the northern terminus of the section (LSD: W of 3, T7 R25 S19 1/4 sec. SE quadrant of 8 to SE quad of 16). The co-ordinates of the Gilchrist and Cypress Lake sections are 72F/11,12UXK166851 (LSD: W of 3, T6 R26 S36, 1/4 sec. SW quadrant of 4), and 72F/6,12UXK178918 to 177929 (LSD: W of 3, T6 R27 S12 1/4 sec. NE quadrant of 6 to NW quadrant of 7), respectively. Thin-sections to evaluate the usefulness of micromorphology as a tool for till genesis studies were obtained from the Coriander Section. This is a small section north of the Frenchman channel on the Wood Mountain map sheet (49°15' N 107° 28'W; UTM grid 72G/11,13UCE691.8102.3; LSD: W of 3, T1 R26 S4 1/4 sec. 11), in southwest Saskatchewan.

The diamicton beds sampled for this investigation are typical of the glacigenic diamictons observed in Alberta and Saskatchewan. Despite their diverse origins the grain size distribution of the diamicton beds is commonly composed of 23 to 38% clay, silt and sand with between 5 and 12% phenoclasts. Any important deviations from this distribution pattern are noted in the text.

Description of the thin-sections

Subglacial meltout till, Battle and Lyons creeks

Twenty-two thin-sections were made from block samples obtained from the subglacial meltout till unit exposed along the Battle and Lyons creeks (Fig. 5.1). The block samples were obtained of (1) the black, platey diamicton that forms the bulk of the unit, (2) the contact zones between the diamicton and enclosed bedrock slabs, (3) the margins and interiors of large bedrock slabs, and (4) diamicton encasing bedrock stringers. In thin-section, the diamicton possessed very low porosity and moderate to well-developed skelsepic and skel-lattisepic fabrics (Fig. 5.2A). Rare diamicton intraclasts were observed. Rounded and elliptical shale phenoclasts with uniform internal birefringence patterns (Fig. 5.2E) make up from 10 to 35 % of the area of the thin-sections. The common size range of these phenoclasts ranged from coarse sand to very fine pebbles. Some of the shale phenoclasts had flat-iron or bullet shapes. Neither diamicton intraclasts nor shale phenoclasts display disruption or intrusion of the surrounding matrix into their interiors.

Thin-sections BC, BC4a and LC8 contain bedrock stringers surrounded by diamicton (Fig. 5.2B, C, D). The boundaries between the stringers and the encasing diamicton are irregular, not sheared. Diamicton extends into the bedrock stringer and the stringer appears to have undergone soft sediment deformation. No unistrial plasma fabrics were observed in association with the thin bedrock stringers. Where clay-rich diamicton has penetrated into the bedrock stringer, the margins of the intrusion are sharp with no intermixing of bedrock and diamicton.

Thin-sections BC9, and LC3, LC6, LC8 contain the contact between a large shale slab and the surrounding diamicton (Fig. 5.2C). The contact between the shale slab and diamicton is distinct but irregular, with no evidence of shearing. The shale portion of thin-section BC9 consists of agglomerated shale pebbles. Clay-rich diamicton lacking grains coarser than fine sand was observed along iron-stained joints in the shale. Under normally polarized light, this diamicton is almost indistinguishable from the shale it has penetrated, but under cross-polarized light the diamicton stringer is easily recognized. In the diamicton portion of the thin-section beneath the shale slab, the diamicton contains fine sand to pebble-sized shale phenoclasts. The boundaries of the fine to medium-grained quartz sand and pebbles are clearly visible. Silt and clay make up about 30 to 60% of the area of the slide. Shale phenoclasts and grains in the diamicton, have uniform birefringence patterns and extinction angles that differ from the surrounding diamicton. The plasma forms a moderately developed skel-lattisepic fabric around pebbles.

Thin-section LC1 (Fig. 5.2C) from the center of a large shale slab (greater than 0.5 m by 1.0 m) contains an agglomeration of spherical and oval sub-units separated by joints whose surfaces are iron-stained. No diamicton is present between the shale sub-units.

Subaerial glacigenic debris-flow complexes

Unit 2, Battle Creek, Saskatchewan.

Only seven thin-sections were obtained from the subaerial glacigenic debris-flow complex that forms the upper unit of the Battle Creek section (BC1). These thin-sections contain 60 to 75% clay and silt. Sand and coarse silt grain boundaries within these thin-sections are indistinct and obscured by the large volume of matrix material (Fig. 5.3A). No plasma fabric was observed (insepic plasma fabric) in the thin-sections (Fig. 5.3A) but irregular diffuse bands were visible in some thin-sections. Shale phenoclasts, (fine-sand to pebble-sized) make up 15 to 20% of the area of the thin section. They have uniform extinctions when rotated under cross-polarized light, with no evidence of disruption and no noticeable preferred orientation of the clay within the pebble.

Cooking Lake moraine thin-sections, central Alberta.

Eight thin-sections were obtained from compact clayey diamicton layers from hummock sections in the Cooking Lake moraine. Many of the thin-sections were texturally banded, and thin stringers of sand and silt, often with loading features, were observed in some of the thin-sections. The thin-sections had abundant matrix (60 to 70%) and few sand grains or phenoclasts. The plasma fabrics are generally insepic but in areas where the clay and silt content was low, poor to moderately-developed skel-lattisepic plasma fabrics are visible. Diamicton intraclasts are present. Some of the thin-sections contained distinct opaque pink bands. (X-ray and chemical analysis showed that the pink coloration was due to an amorphous iron compound that coated the clays and phenoclasts of the diamicton.) No shearing, kinking, streaking or other signs of penetrative deformation were observed in the thin-sections.

Thin-sections from glaciolacustrine sediment complexes

Cypress Lake thin-sections.

The thin-sections analyzed came from Unit 2 of the Cypress Lake section. The unit consists of thin sub-horizontal clayey diamicton beds, 5 to 15 cm thick, separated by laminae and beds of normally graded silt. Rare clay laminae were observed between diamicton beds. Thin sections CL1B, CL1C, CL1D, CL1E, and CL2b possess a low percentage of pore space (3 to 10%), have no visible grain alignments, and insepic or poorly developed skelsepic plasma fabrics. Silt and clay compose between 50 and 70 % of the area of the thin-sections. Grain boundaries are indistinct. Fifteen to twenty percent of area of thin-section CL1B is composed of rounded to oblong shale clasts. These clasts range from very fine pebble to fine sand in size and have uniform internal extinctions. The thin-section is crudely banded with coarse zones separated by zones with greater percentages of silt and clay. CL1C contains a silt band in which irregularly shaped silt intraclasts are visible. Normal grading within the silt band was observed. Loading features deform the upper contact of the silt band. CL1D and CL1E also contain isolated sub-rounded silt intraclasts within diamicton layers. No deformations or disruptions of any of the intraclasts were observed.

Canal section, East Block of the Cypress Hills.

Thin-sections CS1 and CS2 are from the unbedded diamicton at the base of the Canal section. The slides have insepic plasma fabrics and abundant silt and clay that obscures the boundaries of the skeleton grains. Grains are few and shale clasts make up only 1 to 2% of the area of the slides. The thin-sections are uniform in appearance and no banding or laminae or shear structures are visible in the thin-sections.

Glaciolacustrine debris flow complex, Battle River, Alberta.

The glacigenic sediments exposed in rivercuts and gravel pits along and near the Battle River of central Alberta (Fig. 5.1), have been interpreted as a glaciolacustrine debris-flow complex (Kulig 1985). A repetitive sequence of five units is found in the exposures. The units are (from base to top): an unbedded diamicton (Unit 1), diamicton layers intercalated with sand laminae (Unit 2), diamicton layers intercalated with silt laminae (Unit 3), diamicton layers intercalated with silt beds (Unit 4), and silt-clay couplets interbedded with normally graded silt beds (Unit 5). The sequence and the diagnostic sedimentary characteristics of each unit are given in Figure 5.4. Twenty-eight thin-sections were produced from the five units.

Thin-sections from Unit 1 contain diamicton intraclasts and have insepic or poorly developed skelsepic plasma fabrics. The matrix in the thin-section obscures the skeleton grain boundaries. No textural banding or layers were visible.

Thin sand laminae impart a stratified appearance to Unit 2, and are readily visible in the field and thin-section (Fig. 5.4). However, details such as the normal grading cycles only four to five grains thick in these sand laminae are only visible in thin-section (Fig. 5.5C). Abrupt contacts were observed between sand laminae and diamicton interlayers or pebble-sized phenoclasts projecting above the diamicton upper surfaces (Fig. 5.5B, C, D). Diamicton intraclasts, silt and clay intraclasts, and pebble-cored diamicton intraclasts are present in all the thin-sections. These intraclasts range from medium-sand size to medium-pebble in size. Deformation, shearing, or smudging of the soft sediment intraclasts was not observed. Thin-sections from the diamicton layers of Unit 2 have insepic plasma fabrics but small domains with skelsepic plasma fabrics are also present.

In Unit 3, silt laminae separate thin (5 to 10 cm thick) sub-horizontal clayey diamicton layers. Thin-sections reveal diamicton intraclasts, silt intraclasts, and intraclasts composed of silt and clay couplets. These intraclasts have the same size range as mentioned and were present in the diamicton interlayers and the silt laminae. The number of silt and clay couplet intraclasts and silt intraclasts is greater in Unit 3 than in Unit 2. As in Unit 2, no evidence of deformation of the intraclasts (streaking, smearing, disruption) was observed (Fig 5.5e). Normal grading within the silt laminae was also seen. There are sharp erosive contacts between adjacent silt lamina. The diamicton layers usually possess insepic plasma fabrics, but some poorly developed, skelsepic plasma fabric zones are present.

In Unit 4, the volume of the silt interlayers increases substantially compared to Unit 3. Each silt interlayer is formed of several normally-graded silt laminae. Sharp sub-horizontal contacts separate each lamina. Deformation of bedding planes around pebble-sized and smaller phenoclasts (dropstones) and irregularly-shaped diamicton intraclasts, is readily traced. The diamicton beds between the silt interlayers differ from those of Units 2 and 3 since they are formed of intraclasts suspended in a silt matrix. The intraclasts consist of angular, irregularly-shaped silt, and silt and clay couplet intraclasts. The plasma fabrics of these diamicton beds are insepic.

Unit 5 consists of normally graded silt laminae and couplets of silt and clay. Normal grading within the silt laminae and in the silt component of the rhythmite couplets is readily observable (Fig 5.5f). Sharp contacts are observed between successive laminae and silt-clay couplets. Deformations and draping of silt laminae over and beneath irregularly shaped diamicton and silt intraclasts and small pebbles were often seen.

Coriander section, southwest Saskatchewan.

The Coriander section contains an indurated diamicton with sand and silt laminae intercalated between diamicton beds near its base (Fig. 5.6). Thin sections were used to provide an initial assessment of the section. Thin-section W3-8 has less clay in its matrix, contains only 1 to 2% shale pebbles and has a poorly developed skelsepic fabric. W3-1 is texturally banded with coarse and fine layers grading into each other. It has few phenoclasts and only a single shale pebble. The slide is predominantly insepic with some moderately developed skelsepic zones near its top. W3-1 also contains numerous textural bands and layers of very fine sand. This slide has an insepic plasma fabric. No shale phenoclasts are present in W3-1.

Interpretation and discussion of the thin-sections.

The thin-sections from the three glacial environments have some broad differences. The subglacial meltout till thin-sections have moderate to well-developed skelsepic and skel-lattisepic plasma fabric, whereas insepic and poorly developed skelsepic fabric is observed in thin-sections from subaerial glacigenic debris-flow or glaciolacustrine debris-flow complexes. The meltout till also appears to contain less silt and clay, and the boundaries of grains within these thin sections are clearly visible. In contrast, in thin-sections from debris-flow complexes, the matrix volume is higher and the grain boundaries are obscured. Both the debris-flow units and the subglacial meltout tills had a low percentage of pores and were dense and compact. The visual difference in texture differs from the sampled bulk texture and probably reflects incorporation of the diffuse and fine laminae of sand and silt into the bulk samples. In thin-section, these diffuse bands are detectable.

Van der Meer et al. (1992) described a "glacial imprint" in sub-glacial till thin-sections characterized by unsorted sediment with skelsepic plasma fabrics, high density, and a few, unconnected pores. The moderate to well-developed plasma fabric of the meltout tills, and insepic plasma fabric of the resedimented units described in this study agree with Van der Meer et al.'s criteria for identifying subglacial till. But both the meltout till and debris-flow sediments described in this study have high densities and low pore volumes. Additionally, the diamicton beds from all three environments are by definition composed of unsorted sediments with a wide range of grain-sizes, therefore this criterion does not permit differentiation between glacial sediments from subglacial, subaerial and glaciolacustrine environments and may not permit recognition of non-glacial sediments that are compact and contain reworked erratic phenoclasts. Of the four criteria Van der Meer et al. (1992) use to identify subglacial tills, only the strengths of the plasma fabrics appear to separate glacigenic debris flow deposits from subglacial till. Additional criteria are therefore needed to develop a reliable genetic interpretation.

Sand, silt, and clay strata and diffuse banding are commonly observed in the glaciolacustrine and subaerial debris-flow complexes. Shaw (1982) described stratified meltout tills and Muller (1983) described stratified lodgement tills, that could be confused with the stratified subaerial and subaqueous debris-flow complexes. Thin-section analysis may be able to differentiate between these sediment types. Though the sand and silt laminae and beds intercalated with the diamicton beds can be observed and traced in outcrop, only thin-section analysis permits in depth examination of features such as the normal grading within the sand and silt laminae, the nature of the contacts between sorted layers and the diamicton layers over- and underlying them, rip-up clasts in the sand and silt laminae and their internal features, pebble-sized dropstones and their associated structures. The in-depth analysis of the laminae has shown that they are primary depositional features not shear planes. The unit therefore was not formed by lodgement. The suite of features such as the normal grading, the sharp contacts between laminae, the abundance of intraclasts of all types, and the draping structures around phenoclasts (dropstones) could not be produced by meltout. The suite of features could only arise if the diamicton layers and laminae were deposited subaqueously. The normal-grading, sharp contacts between silt laminae and rip-up clasts within the sand laminae prove that the laminae were deposited in a glaciolacustrine setting from underflows (Kulig 1985, chapter 3, this volume). The sharp irregular character of the contacts between the sand laminae and the underlying diamicton layers indicates that the underflows were not of sufficient strength to erode the tops of the diamicton layers flat. Loading of the laminae by overlying diamicton layers indicates that the diamicton layers were deposited episodically. These stratified diamictons could not have originated by meltout (Shaw 1979) or lodgement (Muller 1983) and could only be deposited by an episodic process (debris-flows) that permitted the accumulation of the intercalated laminae. Thin section analysis therefore contributed greatly to the interpretation and improved its reliability.

Intraclasts provide useful information on the origin of a diamicton deposit. Intraclasts, especially the diamicton or pebble-cored diamicton types, are often only detectable in thin-section, because only a small change in texture or internal structure distinguishes them from the surrounding diamicton. In thin-section, small changes in matrix distribution and microscopic partings around the intraclasts allow them to be separated from the encasing diamicton. May (1977) associated silt intraclasts with lacustrine deposition. Shaw (1982) used them to separate glaciolacustrine sediment from meltout till deposits. Ovenshine (1970) linked diamicton intraclasts with ice-rafted deposits, whereas Eyles (1979) associated them with subaerial debris-flow deposits. The presence of intraclasts therefore often indicates remobilization. Thin-sections from meltout till lack soft-sediment intraclasts, whereas the debris-flow deposits contain a variety of them (diamicton intraclasts, pebble-cored diamicton intraclasts, and silt intraclasts). Subaquatic diamicton assemblages commonly contain graded silt and clay couplet intraclasts derived from the erosion or remobilization of previously deposited beds of silt and clay couplets. These are rare in subaerial resedimented units where diamicton intraclasts and pebble-cored diamicton intraclasts predominate.

Debris-flow sediment units can also be over-ridden and re-incorporated into the base of the ice. Examination of the internal structure of intraclasts can separate primary deposits from those that have been over-ridden and thereby subjected to shearing or other types of penetrative deformation. Van der Meer et al. (1985) observed rounded diamicton intraclasts and silt intraclasts within a diamicton that they interpreted to have formed subglacially. In thin-section, the diamicton intraclasts were observed to be penetratively deformed, and unistrial plasma fabric zones and fine-grained shear zones were present. The penetrative deformation of these clasts and the evidence of shearing indicated that the deposit was polygenetic. Van der Meer et al. (1985) determined that pro-glacial debris-flow material and pond sediment containing diamicton and silt intraclasts were over-ridden, deformed and incorporated into basal ice during a readvance. When released, the sediments retained characteristics of the earlier pro-glacial debris-flow origin but were altered by the over-riding. The most important alteration was the shearing and deformation of the diamicton intraclasts. The presence or absence of diamicton and silt and clay intraclasts is therefore not sufficient proof that a diamicton is a glacigenic debris flow. The till, silt and diamicton intraclasts must be analyzed microscopically to determine whether they had undergone penetrative deformation or shearing. At the same time the plasma fabric of the diamicton can be determined. True debris-flow sediment should have insepic or poorly developed plasma fabric while subglacial tills should have skelsepic fabrics and may have pronounced shear zones. The sediment from the Battle River in Alberta has insepic plasma fabric, undeformed diamicton and silt intraclasts and lacks any evidence of shearing or deformation within the diamicton layers. Interpreting them as primary, glacially-derived debris-flows is therefore reasonable.

Micromorphological examination of sedimentary features can also provide information that confirms or elaborates on the genesis of a glacigenic deposit. Micromorphological examination of the contacts between bedrock slabs and the underlying diamicton from the Battle Creek meltout till showed that minor irregularities along the slab edges were preserved. No evidence of shear such as unistrial plasma fabrics, kink folds, or streaking and smudging of the edges of the bedrock slabs was observed. If the blocks and slabs were thrust into position, the soft bedrock phenoclasts should have been sheared and smudged (Marcussen 1975). Deposition by passive meltout, on the other hand, would release the bedrock blocks with their edges intact. Meltout would also account for the diamicton injections into the bedrock blocks along their margins. Melting of buried ice could generate a saturated sediment that if unable to drain, would be unstable and able to flow diapirically. The diamicton observed in thin section along joints of the bedrock slabs could have entered the bedrock slab when it was saturated and plastic. Thin-section analysis therefore eliminates lodgement as the mechanism that formed the lower unit at Battle Creek and indicates that the deposit was saturated at the time of its formation.

Thin-sections also provide clues to the origin of the oval and bullet-shaped shale pebbles observed in all the sections along the Battle and Lyons creeks and at the Cypress Lake sections. Bullet-shaped phenoclasts are most often associated with lodgement tills (Boulton 1976), but can also be observed in meltout tills (Dreimanis 1989). In thin-section, the shale phenoclasts had round or oval cross-sections, uniform internal extinction angles, showed no evidence of penetrative deformation, and had no shear planes, kink folds, or unistrial plasma fabrics surrounding them. The grain-by-grain plastering process, by which a lodgement till is deposited, would deform and smudge the soft shale phenoclasts. A possible mechanism that could form the shale phenoclasts will now be presented.

Thin-sections from the interior and margins of shale blocks show that the shale blocks are highly jointed and are formed of agglomerated rounded and oval-shaped shale subunits. Thin sections BC4, BC9, LC1, and LC2 show diamicton intrusions along joints separating a shale block into smaller pieces in which the agglomerated shale masses are still visible. If the diamicton intrusions were thicker or the block became disaggregated during transport, the rounded and oval shale phenoclasts that form the agglomerations would become separated and encased in a diamicton matrix. The shape of the shale phenoclasts is therefore inherited from the original shale blocks.

The break-up of the shale blocks appears to be a multi-stepped process. The presence of three stages (intact shale blocks with internal agglomerations, shale blocks split along joints by small diamicton injections, and shale phenoclasts encased in diamicton) in a single section indicates that the blocks had undergone multiple stages of erosion, transport, comminution, and redeposition and that the sections are a blend of local and far-traveled material.

The Coriander thin-sections provide a test of the usefulness of micromorphology in till-genesis studies. The thin-sections were obtained from a glacigenic deposit. The insepic plasma fabric, abundance of matrix, and banded nature of the diamicton in thin-section favor but do not prove that the unit is a glacially-derived debris-flow. The thin-sections also do not indicate whether it is a subaerial or subaqueous deposit. Similar difficulties would be met in the Cooking Lake moraine where some of the sediment in the hummocks were deposited in shallow restricted supraglacial ponds. Thin-sections that included only the pond sediments could lead to the conclusion that it was a glaciolacustrine depositional environment and not a subaerial debris-flow setting.

The use of thin-sections in genetic studies is also greatly affected by the preservability of structures within a glacigenic unit. Kulig (1985, 1993) noted that post-depositional diagenesis, especially dewatering, irrevocably destroyed large sedimentary features such as silt laminae (reduced from a bed containing multiple laminae to a thin silt stringer in which individual laminae were unrecognizable) and diamicton pebble intraclasts (reduced from a unit consisting of agglomerated diamicton intraclasts clearly visible in the section to a homogeneous compact diamicton layer in which individual intraclasts were no longer visible even in thin section). Diagenetic alteration of such large features must have had a major impact on the plasma fabric of the unit. Plasma fabrics therefore may also have a post-depositional component that has to be accounted for.

Thin-sections are not a substitute for detailed field investigations in which a suite of sedimentary structures and the total assemblage of features and units can be assessed. Thin-sections are most useful for studying contacts, sand, and silt laminae, and internal features of the diamicton such as shear planes, kink folds, and other features that are difficult or impossible to examine in the field. The proper evaluation of internal sedimentary structures aids in determining the origin of a diamicton and permits a more detailed and accurate interpretation of its genesis. Thin-sections therefore are an important adjunct to field description.

The origin of plasma fabrics

Since the type and strength of the plasma fabrics appear to differ between subglacial and re-sedimented glacial diamictons, the origin of the plasma fabrics is important. Korina and Faustova (1964) stated that the fibrous network (skelsepic plasma fabric?) visible in their till thin-sections was inherited from the ice. Lafaber (1964) attributed formation of skelsepic and lattisepic fabric to rotational movement under pressure. The plasma fabrics therefore may reflect the original structure of the ice or movements within the sediment as it was being released. Subglacial meltout tills form by slow, in situ melting of debris-laden stagnant ice beneath an overlying cover of ice and debris (Shaw 1979, 1982, 1987, Lawson 1979, 1989; Dreimanis 1989). The deposit is usually compact (Lawson 1979; Dreimanis 1989) but if high pore pressure was present the subglacial meltout till can be loose and porous (Ronnert and Mickelson 1992). When the debris is released from the ice it can undergo substantial reorientation due to the generally uneven distribution of debris (Elson 1961, Lawson 1979, 1989). This reorientation and compression beneath a confining mass of debris and ice may involve sufficient rotational movement to form a well-developed skelsepic plasma fabric. This reorientation of debris during meltout makes it unlikely that the debris once released, would retain any of the fine structure it possessed in the ice. The skelsepic plasma fabrics are therefore unlikely to be inherited from the ice as suggested by Korina and Faustova (1964). The plasma fabric observed in a subglacial meltout till may therefore be the result of rotational movement and compression during the early stages of deposition from the melting debris-laden ice. The plasma fabric of a subglacial till differs from that of skelseptic fabrics formed in a soil layer, because in soil layers, only clays and very fine silts are mobile enough to undergo translocation. In a subglacial meltout till, all debris experiences some degree of reorientation.

The formation of the plasma fabric during meltout is complicated by the release of meltwater during meltout till deposition. If the meltwater is unable to drain from the deposit, pore pressure within the sediment may cause small scale internal deformation and reorientation of the subglacially released debris. It is possible that this internal reorientation is what generates the well-developed skelsepic plasma fabrics. Remobilization of subglacial meltout till has been reported by Rusczyñska-Szenajach (1983). The plasma fabrics could also form when meltwater percolating through the deposit causes translocation of clay and fine silt forming coatings around the sand and pebble components. In this second mechanism the deposit probably consolidates as it forms and undergoes little subsequent internal reorientation. The meltwater expulsion could also be accompanied by the removal of the very fine sand, silt, or clay and making the grain boundaries more visible. The fines removed could collect in low pressure zones creating a deposit with that coarsened and fined on a microscale. This may explain the apparently coarser texture of the meltout till in thin-section when compared to thin-sections of debris-flow sediment. Well-developed plasma fabrics therefore could form through several mechanisms or combinations of mechanisms.

The lack of well-developed plasma fabrics in the resedimented deposits may reflect pervasive internal disruption of the plasma fabrics during remobilization that destroyed any internal structure inherited from the ice (if they were inherited from the ice as suggested by Korina and Faustova, 1964). If the plasma fabrics are a result of post-depositional rotational movements under pressure (Lafaber 1964), the debris-flow masses may not have been subjected to sufficient overburden pressures to cause rotational movements and alignment of clay and silt. This may reflect the generally thin nature of the debris-flow deposits studied. Debris-flow sediment at the base of thicker resedimented debris masses may be squeezed sufficiently to cause internal clay realignments and plasma fabric formation. If the translocation of fines by percolating meltwaters is the critical factor then it may be concluded that meltwater percolation did not affect the resedimented deposits as extensively. Lawson (1979, 1989) described Type 1 and 2 debris flows that have low water contents, thin basal-shear zones, and grain sizes that reflect the debris from which they originated. Their low initial water content would reduce the amount of translocation and removal of fines during dewatering. Moreover the low initial water content makes it unlikely that pore-water pressures would be able to attain values high enough to cause internal reorientation. Lawson (1979) also described water saturated Type 3 and 4 debris flows. In these, porewater expulsion removes much of the silt and clay from within the accumulated sediment. The silt and clay are then deposited on the surface of the debris-flow, forming silt and clay caps. The correct combination of the porewater expulsion and overburden pressure could result in the formation of well-developed skelsepic and skel-lattisepic plasma fabrics even in debris-flow deposits.

Although plasma fabric type appears to be an important identifying characteristic, until the mechanism for the formation of skelsepic and skel-lattisepic fabrics is understood it cannot be stated with certainty that subglacial meltout till have skelsepic fabrics and all varieties of debris-flow diamicton do not. No single micromorphological feature should be used to identify a till unit.

Conclusions

Thin-section analysis provides information useful in determining the genesis of a complex glacigenic deposit. Thin-sections, from subglacial meltout till, possess moderate to well-developed skelsepic and skel-lattisepic plasma fabrics and a lower matrix volume. Sand and pebble clasts within the meltout till thin-sections were readily visible and the thin-section had a "cleaner" brighter appearance than those from debris-flow deposits. Glacially derived debris-flow deposits have insepic or poorly developed plasma fabrics. The debris flow sediment is often banded and had a large proportion of silt and clay matrix that obscures grain boundaries making the slides appear darker. Diamicton intraclasts, clay intraclasts, graded-silt intraclasts, and silt and clay couplet intraclasts are present in all the resedimented diamictons. Silt and clay couplet and graded-silt intraclasts are more common in subaquatically deposited stratified glacial diamicton deposits than subaerial debris-flow complexes, but were not observed in diamictons interpreted as meltout tills. The lack of penetrative deformation features within the intraclasts and unistrial plasma fabrics indicate that the glacial deposits investigated were not over-ridden or subject to shearing during deposition.

Micromorphological analysis of sedimentary structures such as contacts, the internal organization of sand and silt laminae, and grading and loading features permits more precise identification of the genesis of a deposit.

In the Lyons and Battle creek sections, thin diapiric intrusions of clayey diamicton into shale blocks along joint surfaces indicate that the diamicton was plastic and able to flow. Shear structures and other evidence of penetrative deformation are not present along the diamicton-bedrock contacts, within the bedrock slabs, or the diamicton that encloses the slabs. The bedrock blocks and slabs could not have been thrust into place. It is more likely that the slabs passively melted out and that the water-saturated diamicton was diapirically injected into the bedrock slabs.

Thin-sections permit more detailed and precise observations to be made of features observable in section as well as on features visible only during microscopic analysis. Thin-sections however are not a substitute for detailed field investigations.

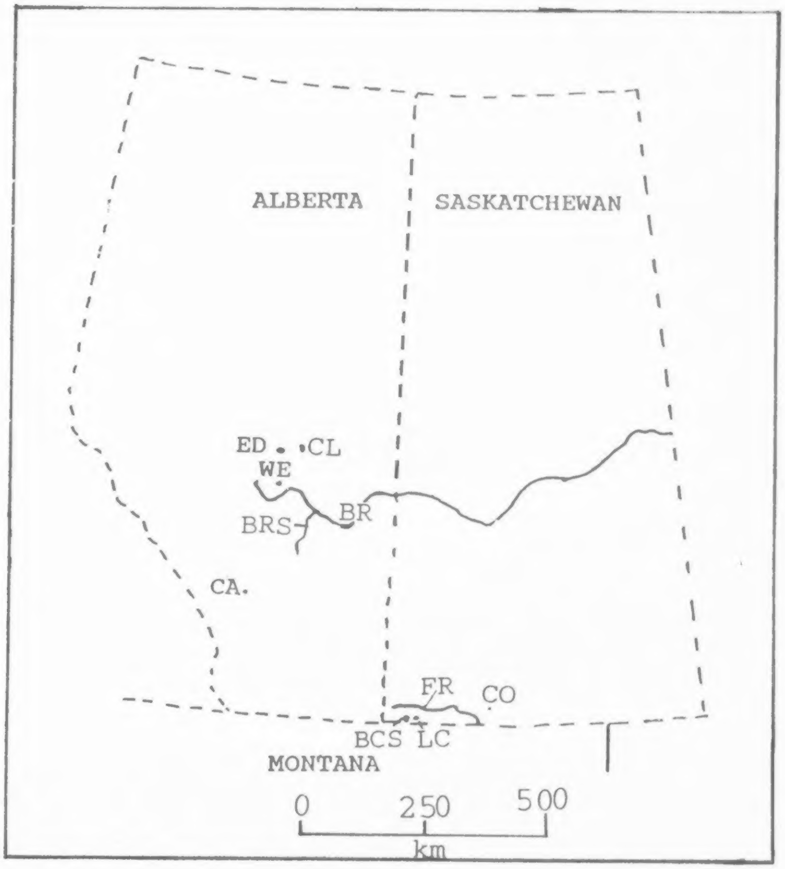
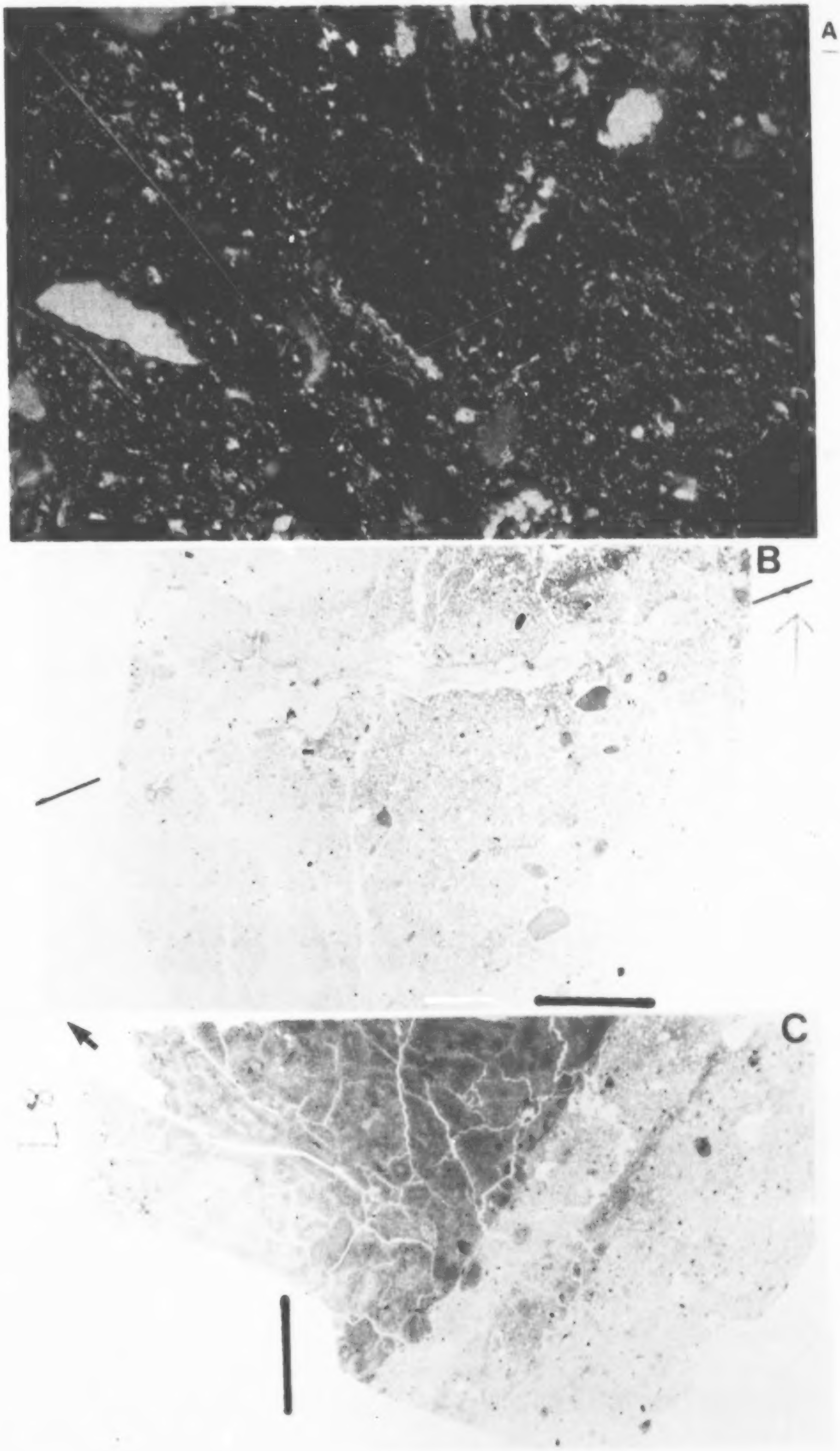

ALBERTA
SASKATCHEWAN
ED
CL
WE
BR
BRS
CA.
FR
CO
BCS
LC
MONTANA
0
250
500
km

Figure 5.1. Location of the sections from which the thin-sections were derived. Letter Code: BCS- Battle Creek section, BRS- Battle River section, BR- Battle River, CA- Calgary, CL- Cooking Lake hummocks, CO- Coriander section, ED- Edmonton, FR- Frenchman River, LC- Lyons Creek sections, WE-Wetaskiwin.

Figure 5.2. Photographs and photomicrographs of thin-sections from the Battle and Lyons creeks sections. 5.2A Photomicrograph of skel-lattisepic plasma fabric in thin-section B4a from the Battle Creek meltout till, magnified 12.5x. Length of the section is 12 mm. Figure 5.2B photograph of thin-section B4a; scale bar is 1 cm. The arrow indicates the vertical plane. Figure 5.2C thin-section LC8 from Lyons Creek 1. The arrow indicates the vertical plane. Scale bar is 1 cm. Note the rounded domains in the shale block at the top of the thin-section and the distinct irregular contact with the underlying diamicton. Fig. 5.2D thin-section BC from Unit 1 at Battle Creek; scale bar is 1 cm. Note the relationship of bedrock stringer to the surrounding diamicton. Fig. 5.2F. Photomicrograph of insepic plasma fabric in the upper debris-flow diamicton of Unit 2 of section BC1. Field of view 12 mm long.


A
B
C

B3
D

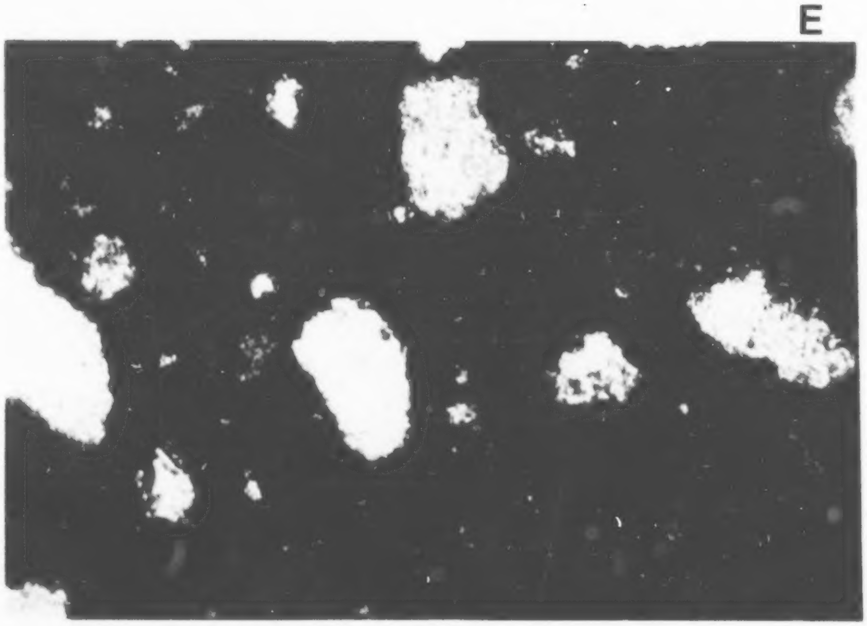

E

Figure 5.3. Intraclast types visible in thin section. Fig. 5.3A thin-section B10 from Unit 1 at section BC 1; magnified 3 x; The photomicrograph shows an undeformed shale clast with clay accumulations around it. Fig. 5.3B Photomicrograph of thin-section NPSF2a, Battle River, Alberta; magnification 12.5 x. Field of view 12 mm. The diamicton intraclast (outlined by white dots) is difficult to distinguish from the surrounding diamicton matrix.

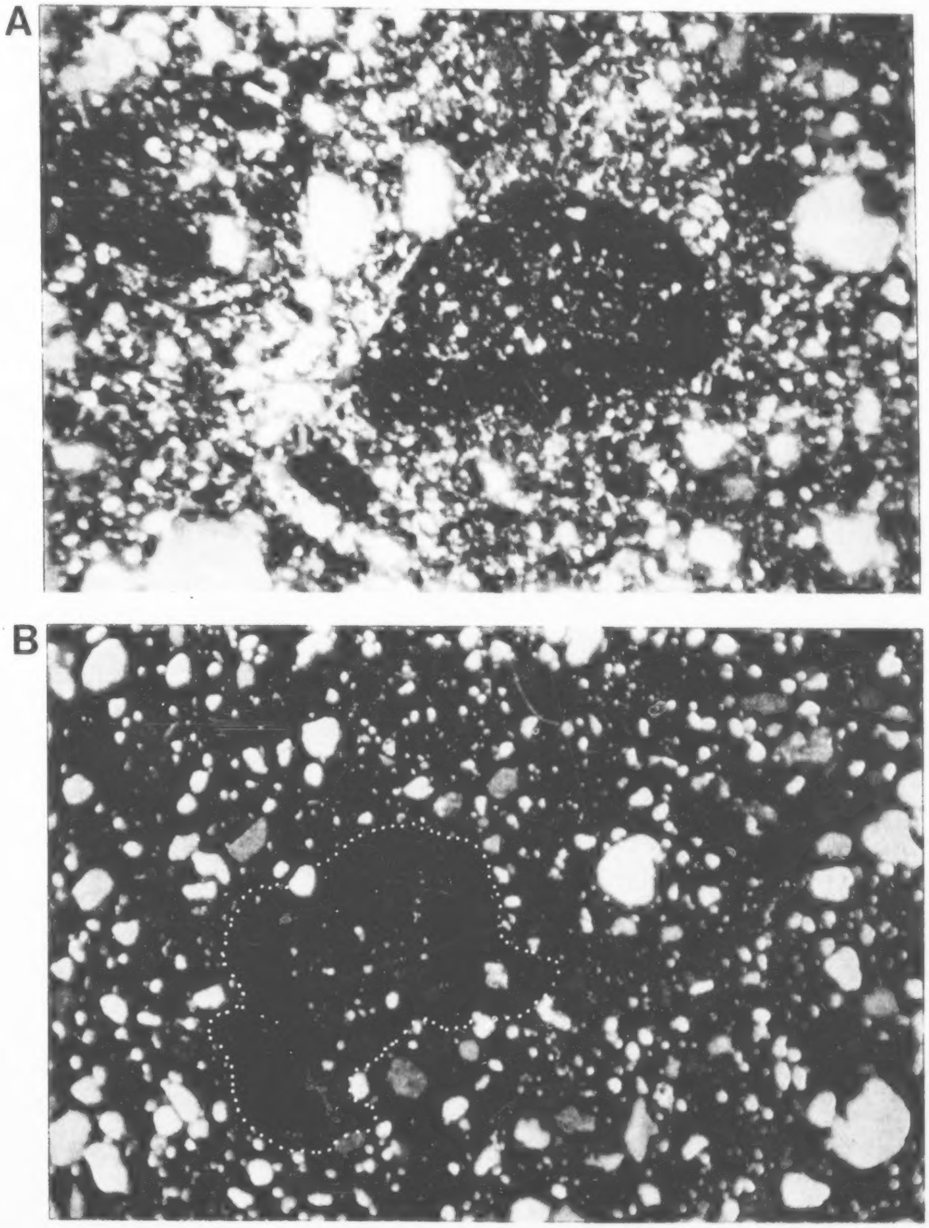
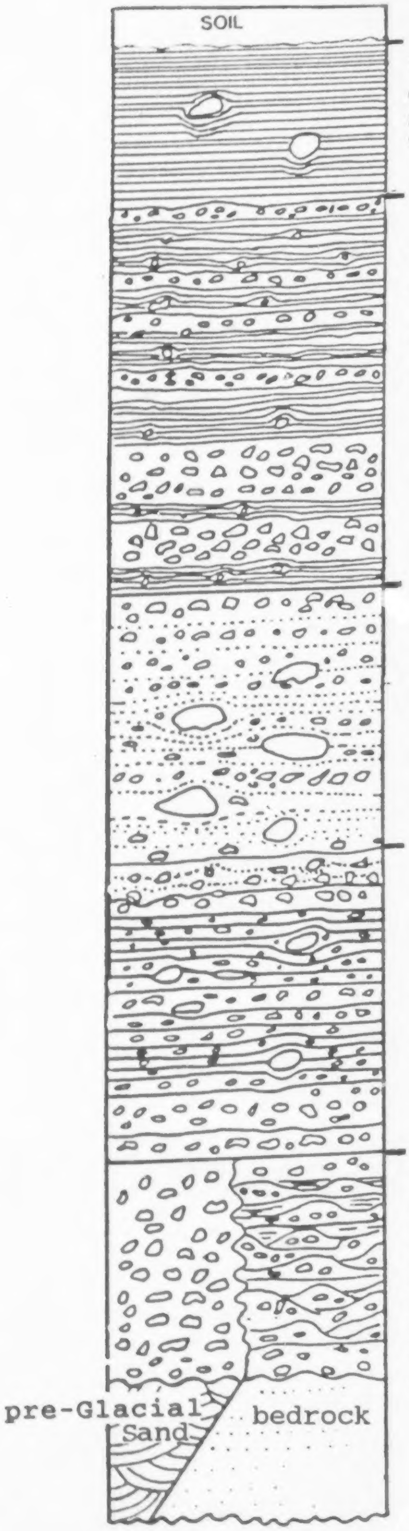


Figure 5.4. The six units in the sections along the Battle River of central Alberta.



Unit 6, Normally-graded silt laminae and silt-clay couplets

Graded silt laminae <2 mm thick, silt and clay couplets 5 to 10 mm thick, numerous dropstones deforming laminae and couplets, gradational contact with Unit 5

Unit 5, Interbedded diamicton and silt layers.

Sub-horizontal diamicton layers 2-17 cm thick, pebbly-textured, abundance of intraclasts (to 80%), irregularly-shaped silt and clay intraclasts, diamicton layers have irregular upper and lower contacts, silt layers formed of 1-10 mm thick normally graded silt laminae, sharp contacts between laminae, rare silt and clay couplets, gradational contacts with Units 4 and 6.

Unit 4, Diamicton layers intercalated with silt laminae.

Sub-horizontal diamicton layers separated by silt lamina and silt layers <1mm thick, Diamicton layers: 2 - 20 cm thick, homogeneous texture, contain silt, clay, diamicton, and pebble-cored diamicton intraclasts, Silt interlayers: 1 - 3 mm thick, rare sand laminae, normally-graded, sharp contacts between lamina, loaded by overlying diamicton layers. Dropstones: singly or in clusters, horizontal a-b axial lanes , deform silt and diamicton layers, abrupt termination of lamina at phenoclasts, phenoclasts draped by subsequent laminae, and diamicton layers. Gradational unit contacts

Unit 3 Diamicton layers intercalated with sand laminae

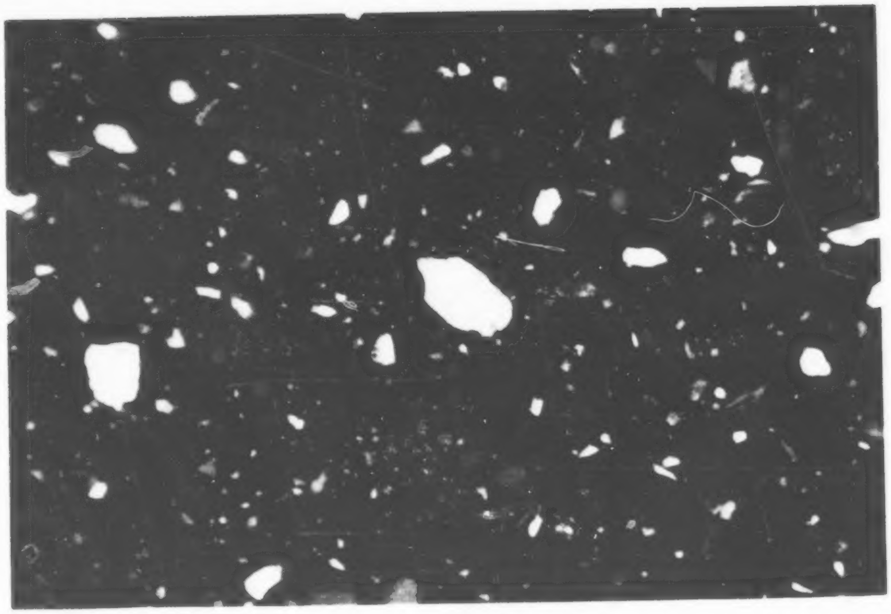
Diamicton: sub-horizontal layers 8 - 200 mm thick, contain diamicton, graded silt and clay, clay, and pebble-cored diamicton intraclasts, random fabrics, homogeneous texture. Sand laminae: 1 grain to 2 mm thick, usually structureless, some normal grading, rare planar beds, pebble lags beneath thicker laminae (>2mm), continuous for 10's of meters. Dropstones: singly or in clusters, predominantly horizontal a-b axial planes, deform sand and diamicton layers, sand laminae terminate abruptly against them, are draped by overlying layers, lee-side shadows and scours.

Unit 1 Unbedded diamicton. Unit 2 Lens-shaped diamicton layers.

Unbedded diamicton 74 to 150 cm thick, diamicton and pebble-cored intraclasts, random fabrics, rare sand and silt stringers, unconformable lower contact with pre-glacial sand and gravel or bedrock, upper contact gradational with unit 3. Lens-shaped diamicton layers, 3 to 6 m in length, 10 to 60 cm thick at center, 1 to 10 cm thick at edges, random fabrics, associated with discontinuous silt beds 10 to 25 cm thick, 2 to 3 m long, deformed by overlying diamicton layers.

Figure 5.5A-F. Photographs and microphotographs of thin-sections from along the Battle River, Alberta. Fig. 5.5A Photomicrograph of thin-section NPW1, showing the insepic fabric of the resedimented diamictons of unit 1 from the Battle River sections. Field of view 12 mm, 12 x magnification. Fig. 5.5B Photograph of JL 3, scale bar is 1 cm. Fine sand laminae and two types of intraclasts present in the stratified diamictons of unit 3. a - denotes diamicton intraclasts, b - is a normally-graded silt -clay couplet intraclast. Fig. 5.5C photograph of NPW9 from unit 3 sand laminae and two types of intraclasts are visible. 1 is pebble-cored intraclasts, 2 are diamicton intraclasts, A fining upwards sequence is visible in the sand laminae. Fig. 5.5D J13, 1.5x, arrow indicates vertical Normal grading visible in rhythmites of unit 6. Fig. 5.5E. Photograph of JL5l. Draping of a phenoclast by sand and diamicton laminae in unit 3. Diamicton intraclasts indicated by 1. 1.5 x magnification, arrow indicates vertical. Scale bar is 1 cm. Fig. 5.5F JL1F 3.25 times magnification, from unit 5 of the Battle River, Alberta sections. Photomicrograph of rhythmite intraclast, the grading within the phenoclast is readily visible. The intraclast shows no evidence of deformation but does have a clay coating around it. 3.25 x magnification, field of view 40 mm.

A



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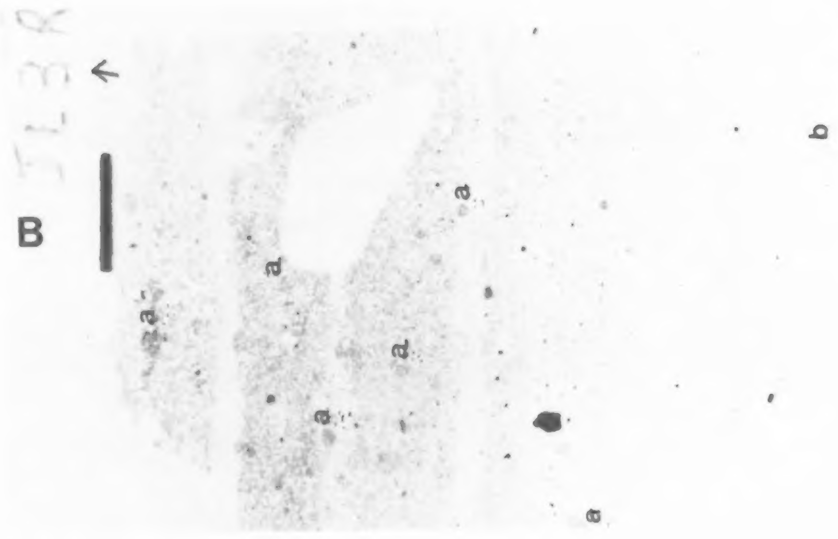
JL3R
a
a
a
a
a
a
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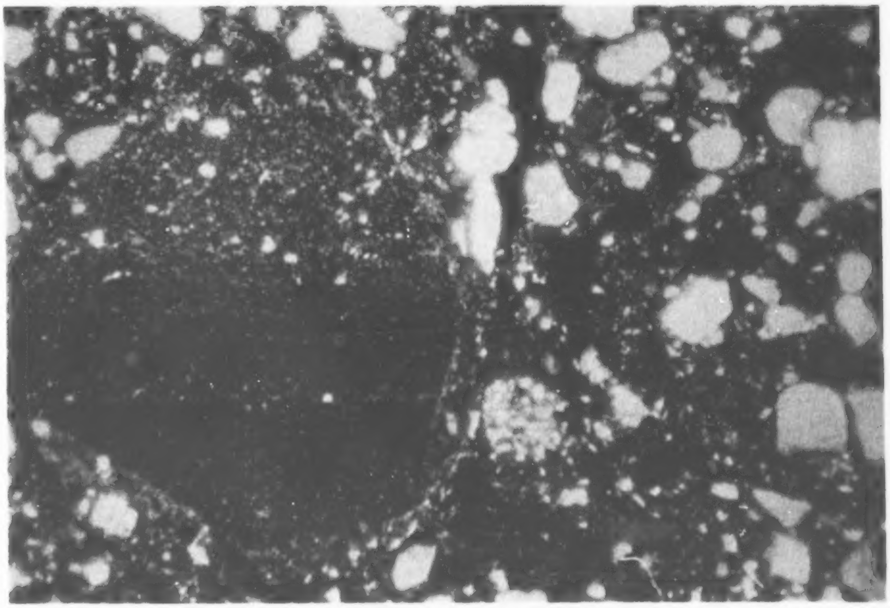
b

F



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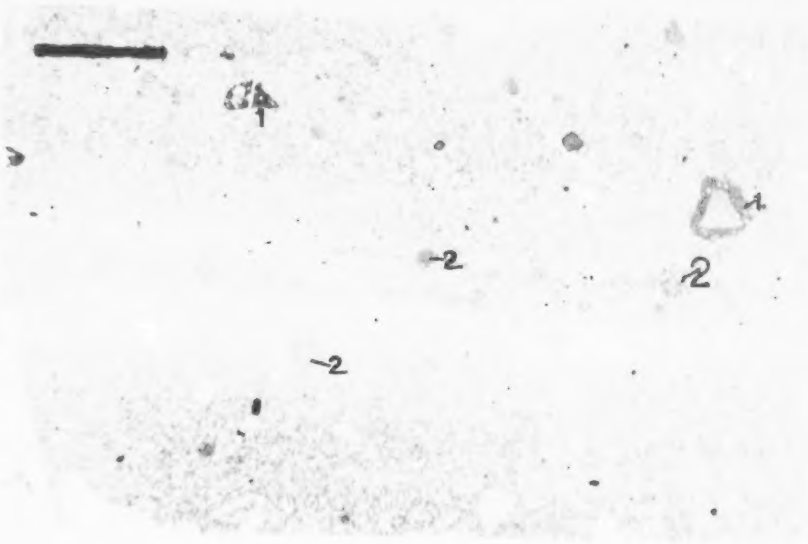


D

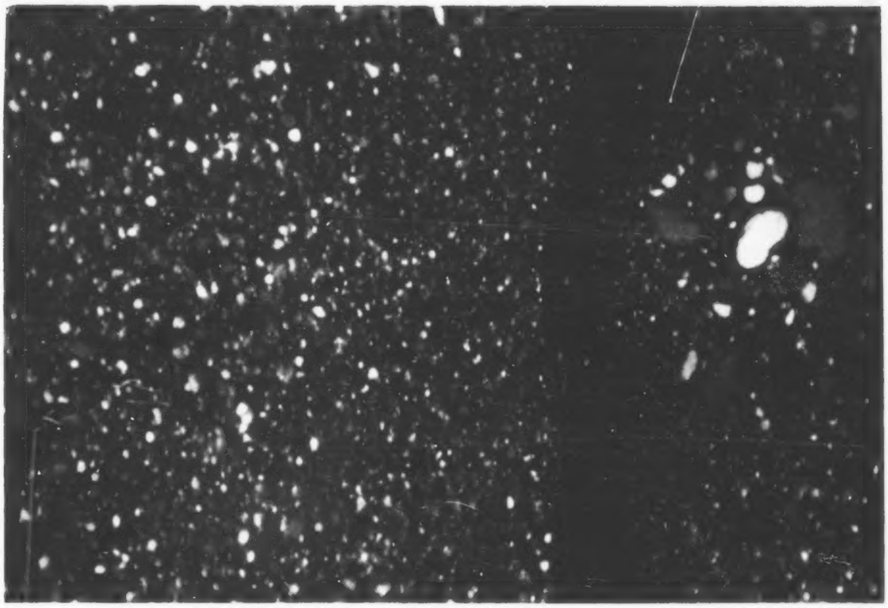
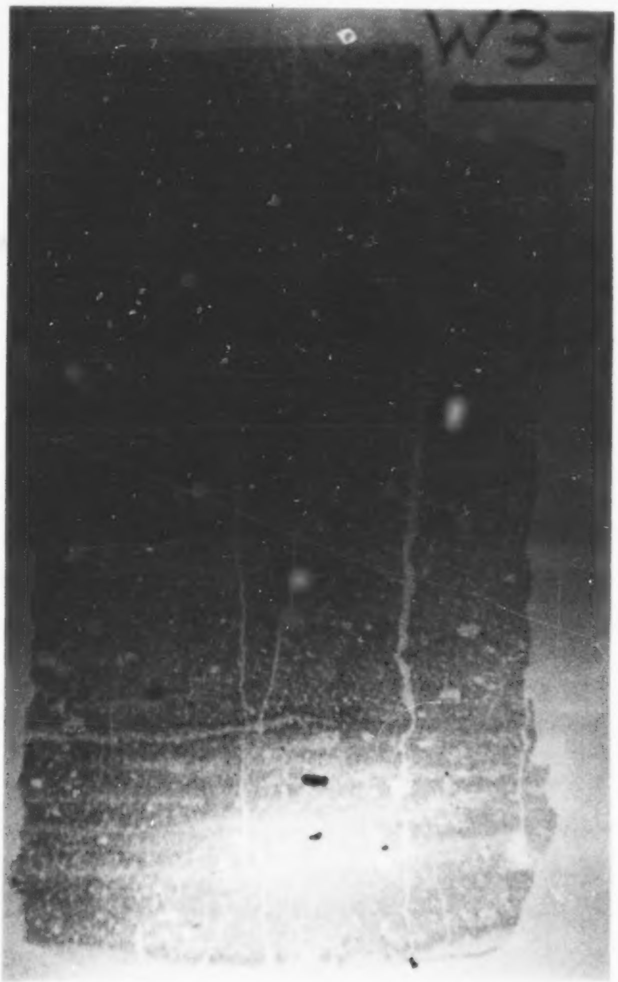


Figure 5.6. Photograph thin-section W3-1 from the Coriander section. The silt and sand banding visible at the bottom of the section. The scale bar is 1 cm.



References cited.

Boulton, G.S., 1978, Boulder shapes and gain size distribution of debris as indicators of transport paths through a glacier and till genesis. Sedimentology, v. 25, p. 773-799.

Brewer, R., 1976, Fabric and mineral analysis of soils. 2nd ed. Wiley, New York, 482 pp.

Brewer, R., and Sleeman, J.R., 1988, Soil structure and fabric: CSIRO Australia, Melbourne Australia, 174pp.

Dreimanis, A.E., 1989, Tills: their genetic terminology and classification, *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam p. 17-84.

Elson, J.A., 1961, The geology of tills, *in* Penner, E. and Butler, J. eds., Proceedings of the 14th Canadian Soil Mechanics Conference, National Research Council of Canada, Associate Committee on Soil and Snow Mechanics Technical Memorandum 69, p. 5-36.

Evenson, E.B., 1970, A method for 3-dimensional microfabric analysis of tills obtained from exposures or cores: Journal of Sedimentary Petrology, v. 40, p. 762-764.

Evenson, E.B., 1971, The relationship of macro- and microfabric of till and the genesis of landforms in Jefferson County, Wisconsin. *in* Goldthwait, R.P. , ed., Till: a symposium, Columbus, Ohio State University Press, p. 345-364.

Eyles, N., 1979, Facies of supraglacial sedimentation on Icelandic and alpine temperate glaciers. Canadian Journal of Earth Sciences, v. 16, p. 1341-1362.

Johnson, M.D., 1983, The origin and microfabric of Lake Superior red clay: Journal of Sedimentary Petrology, v. 53, p. 859-873.

Korina, N.A., and Faustova, M.A., 1964, Microfabric of modern and old moraines: *in* A. Jongerius, ed., Soil Micromorphology. Proceedings of the 2nd working meeting on soil micromorphology. p. 333-338.

Kulig, J.J., 1985, A sedimentation model for the deposition of glacigenic deposits in central Alberta. Unpublished Master's thesis, University of Alberta, Edmonton, Alberta, Canada. 245pp.

Kulig J.J., 1989, The origin of hummocks in the Cooking Lake Moraine, Stop 28. in CANQUA 1989, Late glacial and post-glacial processes and environments in montane and adjacent areas field excursion guide. p. 100-110.

Kulig, J.J., 1993, The preservation og sedimentary structures within glacigenic diamictons: *in* Geological Association/Mineralogical Association of Canada Abstracts with programs 1993, p. A55.

Lafaber, D., 1964, Soil fabric and soil mechanics, *in* Jongerius, A., ed., Soil Micromorphology. Elsevier, Amsterdam, p. 351-360.

Lawson, D.E., 1979, A sedimentological analysis of the western terminus of the Matanuska Glacier, Alaska: U.S. Army Cold Regions Research and Engineering Laboratory. Report 79-9. 122pp.

Lawson, D.E., 1989, Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits: *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits, A.A. Balkema, Rotterdam, p. 147-170.

Marcussen, I., 1975, Distinguishing between lodgement till and flow till in Weischselian deposits. Boreas, v. 4, p. 113-123.

May, R.W., 1977, Facies model for sedimentation in the glaciolacustrine environment: Boreas, v. 6, p. 175-180.

Meer, J.J.M. van der, 1987, Micromorphology of glacial sediments as a tool in distinguishing genetic varieties of till: Geological Survey of Finland, Special Paper 3, p. 77-89.

Meer, J.J.M. van der, and Laban, C., 1990, Micromorphology of some North Sea till samples, a pilot study. Journal of Quaternary Science, v. 5, p. 95-101.

Meer, J.J.M. van der, Rappol, M., and Semeijn, J., 1983, Micromorphological and preliminary X-ray observations on a basal till from Lunteren, The Netherlands: Acta Geologica Hispanica, v. 18, p. 199-205.

Meer, J.J.M. van der, Rappol, M., and Semeijn, J., 1985, Sedimentology and genesis of glacial deposits in the Goudsbourg, Central Netherlands. Medelingen Rijks Geologische Dienst., v. 39, p. 2-29pp.

Meer J. L. M. van der, Rabassa, J.O., and Evenson, E.B, 1992, Micromorphological aspects of glaciolacustrine sediments in northern Patagonia, Argentina: Journal of Quaternary Science, v. 7, p. 31-44.

Menzies, J., and Maltman, A.J., 1992, Microstructures in diamictons - evidence of subglacial bed conditions: Geomorphology: v. 6, p. 27-40.

Muller, E. 1983, Dewatering during lodgement of till: *in* Evenson, E.B., Schluchter, Ch., and Rabassa, J., eds., Tills and related deposits, Balkema, Rotterdam, p. 19-22.

Ostry, R.C., and Deane, R.E., 1963, Microfabric analyses of till: Geological Society of America Bulletin, v. 74, p. 165-168.

Ovenshine, A.T., 1970, Observations of iceberg rafting in Glacier Bay, Alaska and identification of ice-rafted deposits: Geological Society of America Bulletin, v. 81, p. 891-894.

Rappol, M., Haldorsen, S., Jorgensen, P., Meer, J.J.M. van der, and Stoltenberg, H.M.P., 1989, Composition and origin of petrographically-stratified thick till in the northern Netherlands and a Saalian glaciation model for the North Sea basin: Meded. Werkgr. Tert. Kwart. Geologie, v. 26, p. 31-64.

Ronnert, L., and Mickelson, D.M., 1992, High porosity of basal till at Burroughs glacier, southeastern Alaska: Geology, v. 20, p. 849-852.

Rusczyñska-Szenajach, H., 1983, Lodgement tills and syndepositional glaciotectonic processes related to subglacial thermal and hydrologic conditions: *in*: Evenson, E.D., Schluchter, Ch., and Rabassa, J., eds., Tills and related deposits, A.A. Balkema, Rotterdam, p. 113-117.

Shaw, J., 1979, Genesis of Sveg tills and Rogen Moraines of central Sweden: a model of basal meltout: Boreas, v. 8, p. 409-426.

Shaw, J., 1982, Meltout till in the Edmonton area, Alberta, Canada: Canadian Journal of Earth Sciences, v. 19, p. 1548-1569.

Shaw, J., 1987, Glacial sedimentary processes and environmental reconstruction based on lithofacies: Sedimentology, v. 34, p. 103-116.

Sitler, R.F., 1968, Glacial till in oriented thin section: XXIII International Geological Congress. v. 8, p. 283-295.

Sitler, R.F. and Chapman, C.A., 1955, Microfabrics of till from Ohio and Pennsylvania; Journal of Sedimentary Petrology, v. 25, p. 262-269.

Appendix: Terminology

Many of the terms used in this and other papers on the micromorphological features of glacial sediments were developed by soil scientists for description of soil thin-sections. These terms are not often used in geological literature and therefore a brief definition of them is provided here. The definitions are derived from Brewer (1976) and Brewer and Sleeman (1988).

skeleton grains - are individual grains that are relatively stable and immobile (i.e., they are not subject to translocation). They are composed of mineral and other resistant grains that are larger than clay size.

plasma - is the mineral or organic material of clay size that is relatively soluble and is not bound with the skeleton grains. Soil. plasma is readily transported and is therefore mobile.

plasma fabrics - The arrangement of the plasma matrix around the skeleton grains is termed the plasma fabric Many plasma fabrics are identifiable in soil material but only a few of these plasma fabrics are commonly observed in glacigenic sediments unaffected by soil formation processes. If more than a single modifying prefix is used then the prefixes are placed in order of increasing importance, i.e., in skel-lattisepic fabrics alignment of plasma along the skeleton grains is less prominent than the lattice arrangement of the plasma.

asepic fabric - no noticeable zones or patches of plasma alignment within the thin-section or along the skeletal grains.

insepic fabric - isolated patches of oriented birefringent plasma.

lattisepic fabric - the plasma is arranged in a lattice work pattern within the thin section.

skelsepic fabric - the plasma is arranged along the outer boundaries of the skeleton grains highlighting the shape of the grains.

unistrial fabric - plasma arranged in parallel or near parallel birefringent striae in a single direction.

striae - alignments of clay and very fine silt observed within a thin section. The striae appear as thin bright linear arrangements of these components and were first reported in Sitler and Chapman (1955). Sitler (1968) stated that the striae paralleled the ice flow direction.

Chapter 6: The glaciation of the Cypress Hills of Alberta and Saskatchewan and its regional implications.

A version of this paper has been submitted to Quaternary International

Introduction

Establishing the glaciation sequence of the Cypress Hills is crucial to understanding the pattern and deglaciation chronology of southwest Saskatchewan, southeast Alberta and the adjacent glaciated areas of Montana (Fig. 1). The evidence presented here has an impact on two controversial issues: (1) the maximum versus minimum extent of the Late Wisconsinan ice (Stalker (1977, 1980), Stalker and Harrison (1977) and Barendregt (1977) postulated a restricted Late Wisconsinan ice while Westgate (1968, 1972), Christiansen (1979), Fullerton and Colton (1986), Dyke and Prest (1987a,b), Vreeken (1986), and others depicted the Late Wisconsinan ice extending into Montana); and (2) the existence of an ice free corridor between the mountains and Late Wisconsinan ice described in Stalker (1980). In addition to new information from the Cypress Hills area, this paper draws on previously published chronologies and stratigraphies. It revises and re-interprets some of the previous work in light of the new information and ideas developed by the author during fieldwork in the Cypress Hills area. There are still many unresolved issues and difficulties but it is hoped that the paper will spark discussion and re-evaluation of earlier ideas. Evidence will be presented in this paper to extend the Pakowki and Etzikom ice margins (Westgate 1965a, 1968) from the Foremost map sheet into the area north of the Saskatchewan Cypress Hills (Fig. 6.3) and enhance the surficial geology map of Klassen (1991). Two new advances, the Middle Creek and Altawan, will also be described (Fig. 6.3). The Underdahl Advance replaces the Wildhorse Advance (Westgate 1968) and the Green Lake Advance (Westgate 1972) as the maximum advance in the study area. For the Cypress Hills region these changes revise the previously published local (Westgate 1968) and regional deglaciation sequences (Christiansen 1979; Clayton and Moran 1982; Fullerton and Colton 1986; Dyke and Prest 1987a, b; Klassen 1989, 1992). The revised deglaciation sequence suggests that the Late

Wisconsinan ice west of the Cypress Hills was not coupled to the ice mass north and east of the Cypress Hills (supporting the view of Shetsen 1984), since the glaciation sequences for the two ice masses differed significantly. Recognition of an independent Late Wisconsinan ice lobe west of the Cypress Hills provides regionally significant insights into the movements and distribution of the Laurentide Ice Sheet over Alberta during the Late Wisconsinan.

Location and geologic setting

The study area straddles the border between Alberta and Saskatchewan, and includes all the Cypress Hills and the immediately adjacent plains to the north and south (Fig. 6.1, 6.2). The Cypress Hills, a preglacial bedrock plateau rising up to 430 m above the prairie surface (maximum elevation of 1420 m), consist of three blocks: the West Block, which extends 70 km into Alberta; the Center Block; and the East Block (Fig. 6.1, 6.2). The blocks have steep northern margins and gently sloping southern flanks. The east flank of the East Block rises gently from prairie level. Vreeken and Westgate (1992) identified six Miocene tephras on the four erosion surfaces of the Cypress Hills, indicating that much of the relief had been created between 10 Ma and 8.3 Ma. The plains to the south are a mixture of hummocky terrain, eroded till plain, and scabland. The dominant glacial feature of the area is the Frenchman channel, which extends eastward in a sidehill position, along the southern flanks of the Cypress Hills and Wood Mountain Upland. North of the Cypress Hills, ice marginal channels, spillways, and zones of hummocky and ridged moraine are present (Westgate 1968, 1972, Shetsen 1987, Klassen 1991). Near Elkwater, Alberta, Vreeken (1989) described glaciolacustrine sediments deposited in a large glacial lake called glacial Lake Downie. Vreeken (1989) states that the lake drained by 13,510 BP. In the area surrounding Maple Creek, Saskatchewan, glaciolacustrine sediment mantles much of the surface (Klassen 1991, 1992). This sediment was deposited in a large lake impounded between the Cypress Hills and the ice to the north and is younger than Lake Downie. The Etzikom moraine described in this paper is part of the Lethbridge moraine described in Dyke and Prest (1987a).

Stratigraphic framework

Presentation of a revised glacial sequence for the Cypress Hills of Alberta and Saskatchewan, requires discussion of the existing state of the stratigraphy and chronology. The first description of glacial sediments in the area are contained in

McConnel (1885). The existing glacial chronology of the study area was derived predominantly from regional investigations far removed from the study area (Bretz 1943; Horberg 1952; Christiansen 1979; Clayton and Moran 1982; Fullerton and Colton 1986; and Dyke and Prest 1987a, b). Early glacial investigations (Johnston and Wickenden 1931; Bretz 1943; Horberg 1952) produced the first maps of end moraines and glacial lakes of southern Alberta and Saskatchewan that included the study area. Horberg (1952) advocated an extensive Late Wisconsin glaciation, that terminated along the Okotoks moraine southwest of Calgary and extended southwards into Montana (see Figure 14c). Lemke et al.'s (1965) Advance 2 margin is equivalent to Horberg's maximum limit. In the 1960's, glacial stratigraphy was concerned with identifying Early, Middle and Late Wisconsinan deposits as well as attempting to trace Illinoian, Kansan and Nebraskan deposits beyond their type areas. Many of these concepts are now considered to be invalid (see Hallberg, 1986) but the terms and chronologies are still cited and often applied to areas such as the Cypress Hills, creating confusion. Another difficulty confronted when revising the glacial history of the Cypress Hills is the inconsistent nature of older published chronologies. Westgate (1968, 1972) used morphostratigraphic units, Christiansen (1979) and Clayton and Moran (1982) discussed events, Fullerton and Colton (1986) described lithostratigraphic units, Klassen (1992) used an unusual type of morphostratigraphic unit. None of these studies are therefore directly comparable.

This investigation details the glacial events that occurred in the study area. The advances described are events. The sediment deposited during these events would fit best into the allostratigraphic category (North American Stratigraphic Code, 1983). Allostratigraphic units are closely related to the morphostratigraphic units used by Westgate (1968, 1972) when he described the drifts within the Foremost map sheet. The sediment deposited during the advances described in this paper are therefore broadly equivalent to the sediment described in Westgate (1968, 1972). To decrease confusion, Chart 1 presents the most commonly cited glacial chronologies, and attempts to loosely link the different types of units.

Westgate (1965a, 1968, 1972) named the drifts on and around the Alberta Cypress Hills and described the glaciation sequence for the Alberta part of the Cypress Hills. His sequence of drifts from oldest to youngest is; Wild Horse (Green Lake), Pakowki (Robinson), Etzikom, and the Oldman (Walsh). The names in brackets from Westgate (1972) are revisions of Westgate's (1965, 1968) chronology. In addition to changing the names, Westgate (1972) altered the ice margin positions but did not provide reasons for the changes. Barendregt (1977); Christiansen (1979); Clayton and Moran

(1982); and Dyke and Prest (1987a,b) use Westgate's (1968) ice margin positions. In this paper the 1968 boundaries are used to be consistent.

Pre-Late Wisconsinan drift sheets and events

Westgate (1965a, b, 1968, 1972) recognized one morphostratigraphic unit, the Elkwater drift (Fig. 6.3), that was deposited by a pre-Late Wisconsin glaciation. He tentatively correlated this drift with the Advance 1 glacial deposits in Montana (Lemke et al. 1965), and the highest drift surrounding the Del Bonita Upland (Stalker 1962). Both of these drifts were believed to lie beyond the limit of the Late Wisconsin glaciation. Westgate did not trace his drift sheets eastward into Saskatchewan. Barendregt (1977) studied the Pakowki-Pinhorn area immediately west of the Cypress Hills and identified three till units (Wild Horse, Pakowki, and Etzikom). Barendregt (1977) found no evidence of the Elkwater drift but observed a single till sheet (the Wild Horse) south of the Pakowki moraine. He concluded that the Pakowki drift was Illinoian or Early Wisconsinan, while the Etzikom was a Late Wisconsinan deposit.

Christiansen (1979) did not discuss pre-late Wisconsinan events and Clayton and Moran (1982) would only tentatively correlate their North Dakotan Early Wisconsinan margin with a band of hummocky terrain in eastern Montana. No attempt was made to correlate this margin to the Elkwater drift. Fullerton and Colton (1986) recognized three pre- Illinoian and at least one Illinoian till in the subsurface in Montana and correlated their surface Illinoian tills (Heron Park, Marples Point, and Kisler Butte) to Westgate's Elkwater drift (Westgate 1968) and Advance 1 of Lemke et al. (1965). Fulton et al. (1984) recognized a single pre-Late Wisconsinan event (Dunmore Glaciation) in the Medicine Hat, Alberta area. Vreeken (1986) after examining the type area of the Elkwater drift, concluded that the drift did not exist.

Klassen (1992) described two landscape complexes that he believed to be older than Late Wisconsinan (Fig. 1). Only the part of the "bedrock terrain with residual drift landscape complex", on the south slopes of the Cypress Hills of Saskatchewan, is contained within the main study area of this paper (see box Figure 1). The part south of the Wood Mountain Upland (Fig. 1) is outside this paper's main study area. This complex has been interpreted by Klassen (1992) as Nebraskan or Kansan. This surface unit consists of isolated granite and carbonate erratics lying on top of the preglacial Cypress Hills Formation. The "bedrock terrain and drift landscape complex" (Fig. 1), a patchy veneer of glacial sediment and till, was interpreted as Early Wisconsinan or Illinoian. This complex therefore could be broadly equivalent to Westgate's (1968, 1972) Elkwater drift (Fig. 6.3).

Late Wisconsinan drift sheets and events

Much of the debate concerning pre-Late Wisconsinan deposits revolves around the location of the Late Wisconsinan margin. Horberg (1952) and Westgate (1968) concluded that the Late Wisconsin glaciation had extended to the Okotoks moraine and south into Montana, where it reached the Advance 2 position of Lemke et al. (1965). Barendregt (1977), Stalker (1977, 1980), and Stalker and Harrison (1977) placed the margin of the Late Wisconsin glaciation in Alberta at the Etzikom moraine (Fig. 6.3) north and west of the Cypress Hills, thereby making the surface south of this moraine pre-late Wisconsinan. Other investigators (Clayton and Moran 1982; Shetson 1984, 1987; Fenton 1984; Fullerton and Colton 1986; Fullerton 1989, unpublished correlation chart; Vreeken 1986) disputed the validity of this margin.

Christiansen's (1979) discussion of Late Wisconsin glaciation in southern Saskatchewan described only two ice positions, a maximum south of the Cypress Hills, and a retreatal position north of the hills. During the Late Wisconsinan maximum (about 20,000 BP), the ice was separated into a northeast lobe covering the Swift Current plateau and a southwest lobe overlying the terrain south and west of Eastend, Saskatchewan (Fig. 6.5, ice margin S1+S2). The two lobes were separated by an interlobate Frenchman channel that was gradually deepened by the meltwater it carried. By Time 2 (about 16,500 BP), the ice margin had permanently retreated north of the Cypress Hills to the Etzikom moraine (Westgate 1968) in Alberta. Christiansen (1979) extended the Etzikom margin eastward north of the Cypress Hills, to the Pelletier Channel (margin S5 in Fig. 5 and shown in Fig. 14f).

Clayton and Moran (1982) attempted to trace Late Wisconsinan dated ice-marginal positions from North Dakota and Manitoba throughout the plains of Canada and Montana. Clayton and Moran's (1982) Phase D margin at the Late Wisconsinan maximum (about 20,000 BP) mirrored the Time 1 ice lobe pattern of Christiansen (1979), (Fig. 6.5, ice margin S1+S3) and was tentatively correlated to the Late Wisconsinan glacial maximum limits in Montana of Lemke et al. (1965). Following an extensive retreat, the ice readvanced to their Phase E position (about 18,000 to 15,000 BP), which closely approximates their Phase D margin (Fig. 6.5, ice margin S1+S4). By Phase F (about 14,000 to 13,500 BP), the ice margin had retreated to the Pakowki moraine (Fig. 6.3, 6.12, 6.14e) in Alberta and extended north of the Cypress Hills and Wood Mountain Upland before entering northeast Montana. After this, the ice receded permanently from Montana. During the next minor advance (Phase I about 12,300 BP), the ice reached the Etzikom moraine but was confined north of the Wood Mountain Upland.

Only the first two stages of Dyke and Prest's chronology (Dyke and Prest, 1987b) affected the study area. During their 18,000 BP maximum, three ice masses affected the Cypress Hills. One ice mass that flowed around the west flank of the Cypress Hills, one that flowed around the east flanks of the Cypress Hills, and a center ice mass that flowed over the Cypress Hills (in the area of the Gap Lobe of this paper. All three ice masses extended southward into Montana). By 14,000 BP, the time of their first recessional margin, the ice had receded north of the Cypress Hills to the approximate position of Christiansen's (1979) Time 2 boundary (Fig. 6.5, ice margin S5). Klassen's (1989) chronology did not show ice-lobe coalescence and did not explain whether his 18,000 BP margin is the maximum or a recessional position. His chronology also placed the ice near Christiansen's (1979) Time 2 margin by 14,000 BP.

Fullerton and Colton (1986) provided an extensive treatment on the surface distribution of the Late Wisconsinan till sheets and possible correlations to the surface tills in Alberta and Saskatchewan. Unfortunately no details about the ice-retreat pattern or ice marginal positions in Montana during the Late Wisconsin glaciation were given.

Vreeken (1986) interpreted evidence from soils and surface units to indicate that the area west and south of the Cypress Hills in Alberta was covered by Late Wisconsinan ice. He placed the Late Wisconsinan margin in the Elkwater area at the Green Lake margin position (the Green Lake margin (Westgate 1972) matches Westgate's (1968) Wild Horse position in this area). Vreeken (1989) dated the Etzikom moraine (part of the Lethbridge moraine) to be about 13,510 BP.

Klassen (1991, 1992) and Klassen and Vreeken (1985, 1987) described the distribution of surface sediments in the Cypress Lake and Wood Mountain map sheets of southwest Saskatchewan. Klassen (1992) speculated on the age and distribution of Late Wisconsinan landscape complexes in the Cypress Lake and the adjoining Wood Mountain map sheets. Klassen (1992) described a "first advance landscape complex" (Fig. 6.1) that consists of hummocky terrain and till plain between the Frenchman channel and an end moraine northwest of the area. Klassen (1992) stated that Late Wisconsinan ice from the northwest did not cover the area and proposed that an earlier Late Wisconsinan advance from the southwest deposited this complex.

A non-stratigraphic study that is particularly relevant is Shetsen's (1984) pebble-lithology investigation. The study identified three ice lobes over Alberta during the Late Wisconsin glaciation (Fig. 6.4). Shetsen's East Lobe, north of the Cypress Hills, was connected to the main Laurentide ice mass to the east, whereas Shetsen (1984) stated that the Central Lobe was connected to an ice divide in the Keewatin District of the Northwest Territories. This lobe extended in a southwesterly direction along the west flank of the

Cypress Hills and terminated in Montana. Shetsen's East Lobe should correlate with the East Lobe of this paper and the main ice mass north of the Cypress Hills. Her Central Lobe corresponds to the West Lobe of this paper. The location of the suture zone between these two lobes is uncertain.

The glacial history of the Cypress Hills region is complex and confusing. The following sections will attempt to clarify the ages of the surfaces, the extent of moraines and try to reconcile as much as possible the conflicting margins and correlations.

Late Wisconsinan glaciation of the Cypress Hills and adjacent plains.

The distribution of the three ice lobes surrounding the Cypress Hills during the Late Wisconsin glaciation is described in this section. A schematic time-distance diagram (Fig. 6.6) illustrates the relationship between the lobes. The advance and retreat sequence in the study area are based on interpretation of glacial deposits, landforms, and drainage networks observed in air photographs, supplemented by field investigations and section stratigraphy. Tills, glacigenic diamicton, glaciolacustrine, and glaciofluvial deposits were studied. Landforms such as moraines, hummocky terrain belts, meltwater channels and spillways, and scablands were also examined. The paucity of dateable materials, prevents establishing an exact chronology, but the relative chronology developed, provides a detailed picture of the area's glacial history. This in turn impacts on the regional deglaciation chronology.

New field evidence (Chapters 2, 3, 4 this volume; Klassen 1992) indicates that the distribution of the lobes differed significantly from that portrayed in earlier reconstructions. This revised ice distribution places the Late Wisconsinan ice lobes farther (higher elevation) onto the southern flanks of the Cypress Hills (Fig. 6.7) than shown by Christiansen (1979), Clayton and Moran (1982) (Fig. 6.5) but the ice was less extensive than shown in Dyke and Prest (1987b). When the three ice lobes extended onto the Cypress Hills, they coalesced on the south side of the hills, and impounded meltwater draining across the Cypress Hills, forming ice-dammed lakes on the south side of the Cypress Hills. Ice rafting of debris across these lakes dropped erratics beyond the limits of the ice lobes. The Elkwater drift in Alberta (Westgate 1968, 1972), (Fig. 6.3), and the "bedrock terrain with residual drift landscape complex" in Saskatchewan (Klassen 1992), (Fig. 6.1), previously interpreted as pre-Late Wisconsinan relict deposits, are reinterpreted here as Late Wisconsinan ice-rafted deposits.

The Underdahl Advance is named for Underdahl Creek, on the southern flank of the West Block, near the limit of the advance (Figs. 6.2, 6.8). The deposits of the Underdahl Advance mark the maximum limit of the Late Wisconsin glaciation in the Cypress Hills. The Middle Creek Advance is named for the Middle Creek channel near its limit (Fig. 6.2, 6.10). The Altawan Advance (Fig. 6.3, 6.11) is named for the Altawan Reservoir near the limit of that advance (Fig. 6.2). The Pakowki and Etzikom Advances (Figs. 6.12, 6.13 respectively), (Westgate 1968), are retained since these advances north of the Cypress Hills in Saskatchewan are continuations of the Pakowki and Etzikom margins in Alberta (Westgate 1968) that have been used in many of the regional reconstructions.

The new glacial sequence of the study area was divided into a series of unequal time periods chosen to highlight the major Late Wisconsinan events. These time slices are illustrated in Figures 6.7 to 6.13: Time I is the Late Wisconsinan maximum; Time IIa illustrates the incision of the Frenchman channel during the earliest stages of retreat ; Time IIb illustrates the formation of the early Middle Creek Channel during continued wastage of the ice; Time III shows the Middle Creek Advance that partially infilled the early Middle Creek channel; Time IV illustrates the limited Altawan Advance; During Time V (Pakowki Advance) the ice retreated permanently north of the Cypress Hills. By Time VI (Etzikom Advance) the ice occupied a position in the extreme north of the study area. Regional positions of the ice margin for the same time intervals are given in Figure 6.14A-F.

Time I. The Late Wisconsinan maximum. About 20,000 BP

The Underdahl Advance

During this advance the ice wrapped around the Cypress Hills and was split into the West, Gap and East lobes, (Fig. 6.7). The West Lobe flowed south and eastwards around the West Block of the Cypress Hills as indicated by westward dipping bedrock slabs and phenoclast fabrics measured in sections along the Battle and Lyons Creeks (Chapter 2, this volume). It terminated in Montana where it reached a maximum elevation of 1650 m (Barendregt, 1977). Oolitic ironstone erratics from east of Yellowknife in the Northwest Territories (Klassen 1989) are found only in areas covered by the West Lobe. Carbonate and granite erratics record the northeastward advance of the West Lobe onto the southern flanks of the West Block (Westgate 1968; Klassen 1989, 1992; Chapter 2, this volume). The erratics are found to an elevation of 1350 m on the north side of the West Block (Westgate 1968), but only 1250 m elevation on the south side of the West

Block. The West Lobe terminated in the area southwest of the Old Man On His Back and Boundary Plateaus (Fig. 6.7).

The ice mass north of the Cypress Hills contains the Gap Lobe and the East Lobe. The East Lobe flowed south and southeastwards across the Swift Current plateau. South of the East Block, the ice continued the south and southeast but part also flowed westwards onto the south slopes of the East and Center blocks (Fig. 6.7, 6.8, 6.14a, b). The East Lobe also terminated in Montana (Fig. 6.7, 6.14a). The Gap Lobe flowed southwards along the pre-existing Gap Creek valley in the Center Block and terminated in a moraine on the south side of the Center Block (Fig. 6.7).

The suture zone between the East Lobe and the West Lobe Hills south of the Cypress Hills is uncertain. Bedrock thrust by ice moving from the northeast is observed on the northeast sides of the Old Man On His Back and Boundary Plateaus and along Highway 18 between Divide and Claydon. Northeast dipping ice-thrust bedrock is also visible in a kaolinite pit 3 km south of Eastend, Sask. South of Frontier, Saskatchewan, near the Montana Border, are eskers deposited by meltwaters flowing from the northeast. These features indicate that the East Lobe flowed south, across the Frenchman Channel to zone along the east sides of the Boundary and Old Man On His Back plateaus. As stated erratics with a western source are restricted to the area southwest of these plateaus in the area covered by the West Lobe. Klassens and Vreeken (1987, Fig. 12.4, p. 115) indicates that the West Lobe had a lower carbonate content than the East Lobe. The coalescence zone between the two lobes as illustrated in Figures 6.3, 6.7, and 6.8 is broad but was in this vicinity. That the two lobes coalesced is shown by the glaciolacustrine sediment in sidehill positions on the south slopes of the East and Center blocks. If the ice had not coalesced, the lakes would not have been impounded on the south flanks of the Cypress Hills (Fig. 6.8) and meltwater would have drained southwards along the gap between the East and West lobes.

The coalescence zone between the East and West lobes north of the Cypress Hills is also problematic. Shetsen (1984) on the basis of pebble lithology differences, placed the suture zone west of the West Block (Fig. 6.4), whereas Klassen (1992) believed it is located near the Gap Lobe. More investigation into the suture zone is needed to determine which viewpoint is correct.

In the Elkwater area Vreeken (1986) placed the Late Wisconsinan limit at the Wild Horse moraine. It is possible that the limit was slightly further to the south, nearer the Elkwater Channel, but incision of the Elkwater Channel obscured this margin. On the north flank of the West Block (Fig. 6.7, 6.9), near the Alberta - Saskatchewan border, a prominent moraine formed along the terminus of the ice sheet during this advance. On the north side of the Center and East blocks, discontinuous ice-pushed ridges, composed of preglacial gravel with less than 1% erratic content, mark the limit (Fig. 6.9). South of the Cypress Hills a terminal moraine across Merryflat between the West and Center blocks marks the limit of the West Lobe there (Fig. 6.8, 6.9). Discontinuous meltwater channels on the north bank of Frenchman channel above Cypress Lake indicate that initially the ice-marginal drainage was confined to ice-walled, or perhaps supraglacial channels.

Surficial deposits left by this advance consist of diamicton and glaciofluvial and glaciolacustrine sediments. Interbedded till and lake sediment (map unit Mx Klassen 1991) were deposited on and near the south flanks of the Saskatchewan Cypress Hills. Sediment derived from the West Lobe was carbonate rich while that derived from the East Lobe was not (Klassen and Vreeken 1987). During the course of fieldwork in the area the author observed that glacial diamictons deposited from the West Lobe contained 10 to 15% shale clasts while diamictons derived from the East Lobe contained few to no shale clasts. Deposition of the thick meltout till observed in the Battle and Lyons creek sections along the Montana border began during this advance. The sections contain 8 to 12 m of basal meltout till with stacked slabs of shale and sandstone overlain by 2 to 8 m of intermixed glaciofluvial and glacially-derived debris-flow sediment. They indicate that the base of the ice had large debris concentrations.

In Montana, the Fort Assiniboine, Loring, and Crazy Horse tills were deposited from the Late Wisconsinan ice (Fullerton and Colton 1986; Fullerton, pers. com., 1991). The Fort Assiniboine Till covers Montana west of the Boundary Plateau, the Crazy Horse Till covers the eastern part of Montana, and the Loring Till covers central Montana from the Boundary Plateau east to the Flaxville Plain (the unglaciated southern extension of the Wood Mountain Upland). The Loring Till is contiguous with the surface Wymark Till (Christiansen 1968), that covers southern Saskatchewan (Fullerton and Colton 1986). Glacial sediment exposed along the Battle and Lyons creeks near the Montana border should be equivalent to the Fort Assiniboine Till. It is unclear whether deposits of the East Lobe can be linked to the Loring Till.

The Underdahl Advance on the southern flanks of the West block in Alberta was at or very near the Wild Horse Advance limit (Westgate 1968). Along the southern flanks of the Cypress Hills in Saskatchewan, it was more extensive than Phase 1 of Christiansen (1979) or Phase D of Clayton and Moran (1982), (Fig. 6.5) but less extensive than shown by Dyke and Prest (1987a).

Summary

At the Late Wisconsinan maximum, the Cypress Hills split the ice into three discrete lobes (West, Gap, and East lobes), (Fig. 6.7). The lobes extended around the hills and ultimately, the West and East lobes reached Montana. Coalescence of the three lobes near Cypress Lake set up the pre-conditions needed for formation of ice-dammed lakes, ice-rafting erratics across the lakes, and incision of the Frenchman channel.

Time II. Wastage from the Underdahl limit and the catastrophic incision of the Frenchman channel, about 20,000 to 18,000 BP.

Coalescence of the three lobes dammed the meltwater that flowed from the northern ice mass across the Cypress Hills. After coalescence and during the earliest stages of retreat, the meltwater initially collected in the lows between the West and Center blocks and the Center and East blocks and then expanded along the southern flanks of the hills. Lakes Graburn, Belanger, and Blacktail were formed, (Fig. 6.8). Scattered erratics perched upon pre-glacial gravels are observed beyond the limit of the Underdahl Advance. These erratics were previously interpreted (Westgate 1968; Klassen 1992) to be remnants of an earlier more extensive ice sheet. But as their distribution matches that of the location of the lakes impounded on the south flanks of the Cypress Hills they have been reinterpreted here as Late Wisconsinan ice rafted materials. Vreeken (1986) also interpreted these isolated erratics as ice-rafted debris.

As the impounded lakes were forming, meltwater flowed south and eastwards along the margins of the ice lobes in small channels that were supraglacial or partially ice-walled (shown by discontinuous glaciofluvial channels on the south flanks of the Frenchman channel). Meltwater drainage past the East Block (from Eastend to the southwest) was probably also carried predominantly in supraglacial or ice-walled channels. As the retreat continued, water accumulated in two additional areas: the coalescence zone of the three lobes, forming Lake Cypress along the south flank of the Center Block, and in Lake Robsart on the plain south and east of Cypress Lake (Fig. 6.8). It is possible that lakes Cypress and Robsart were connected. An additional lake, Lake Fairwell may also have been present on the East Block. Intermixed glacigenic diamicton sediment and lake sediment were deposited in the lakes. The Frenchman channel formed when continued retreat caused the ice dam and moraine impounding Lake Graburn at Merryflat to catastrophically fail. A channel was rapidly cut from Merryflat to Cypress Lake in the thin crevassed, easily eroded ice where the ice ascended onto the southern slopes (Fig. 6.8). The outburst overfilled Lake Cypress and continued eastwards triggering the catastrophic drainage of lakes Belanger, Fairwell(?), and Robsart (Chapter 4, this study). The catastrophic release of Lake Robsart at this time rapidly incised Palisades Coulee which connects to the Frenchman channel (Chapter 4, this volume).

At Eastend, where the Frenchman channel turns southeast, the outburst flowed through either a pre-existing supraglacial channel or followed the surface topography of the ice until it again attained a sidehill position along the Wood Mountain Upland. The channel followed the Wood Mountain Upland southwards into Montana where it joined the Milk River channel (Fig. 6.1, 6.14B). The sidehill positioning of the channel along the Cypress Hills and Wood Mountain Upland required confining ice to the south.

An anastomosing channel network on the Swift Current Plateau in the Shaunavon area was identified by Vreeken (1991) as the product of a subglacial meltwater flow which entered the Frenchman Channel. The timing of this sub-glacial flow is uncertain. Outside the study area east of Val Marie, on the Wood Mountain Upland a set of channels, extend south of the Frenchman channel and are cut by the Frenchman channel. The Frenchman channel therefore, postdates the formation of these channels. These channels may have been cut by protracted meltwater drainage across the upland and may indicate that the ice there was relatively immobile for an extended time. After the Frenchman channel was cut, the ice wasted back from the area south of Cypress Hills. Around Palisades Coulee, the ice stagnated and melted, creating a large tract of hummocky terrain (Fig. 6.9).

The deposition of the subglacial meltout till at the Battle and Lyons Creek which had begun during Time 1 continued. Deposition of the subaerial debris-flow diamicton (Unit 2), described at the Lyons and Battle Creek sections (Chapter 2, this study), most likely occurred at this time.

The active margin receded north and west of Middle Creek and the Altawan Reservoir area. This allowed meltwater draining across the West Block to cut the Middle Creek channel southward from the Merryflat area (Fig. 6.9). The ice wasted back to an unknown position north and west of the Cypress Hills. The wastage of the ice was terminated by the Middle Creek Advance.

The pattern of retreat of the East Lobe and ice north of the Cypress Hills has not been established. Clayton and Moran (1982) stated that after reaching its maximum, the Late Wisconsinan ice retreated from its maximum limit. While there is evidence to indicate that the West Lobe did waste back, it is unclear whether the East Lobe also retreated. The lack of evidence of an extensive retreat may indicate that its movements were independent of the West Lobe and that the East Lobe may have remained relatively immobile.

Summary

During the earliest stages of retreat, meltwater was impounded on the southern flanks of the Cypress Hills. Erratics were rafted across the lakes. Initially these lakes drained south and eastwards along the margins of coalesced ice lobes in small ice-walled channels. Continued retreat expanded the lakes and caused the ice dam impounding Lake Graburn to catastrophically fail. The meltwater followed the break-in-slope eastward along the southern flanks of the West and Center blocks, and overfilled Lake Cypress that failed. This in turn triggered the release of lakes Belanger, Robsart, and Oxtail. Drainage of Lake Robsart formed the Palisades Coulee, that was then surrounded by an isolated mass of stagnant ice. This stagnant ice mass melted, leaving a large tract of hummocky terrain. Eskers which feed into the Palisades Coulee flowed in subglacial and ice-walled channels. These eskers transported the meltwater released from the melting of the stagnant ice in the area. East of Eastend, the water probably followed the course of a pre-existing supraglacial channel. After incision of the Frenchman channel, the West Lobe wasted back to an unknown position. During this time, meltwater drained southwards, along the Middle Creek channel (Fig. 6.9). The movements of the East Lobe at this time have not been determined and it is uncertain whether the East Lobe retreated synchronously with the West Lobe or whether the movements of the two lobes were unconnected.

Time III. The Middle Creek Advance, About 15,000 BP

During the Middle Creek Advance, the West Lobe reoccupied much of the area south and west of the West Block and reached an elevation of about 975 m on the south side of the West Block of the Cypress Hills (Fig. 6.10). Surficial sediments deposited during this advance are varied. Along the west flank of the West block of the Cypress Hills only a sparse cover of surficial sediment mantles the bedrock, but numerous eskers are seen in the channels that dissected this area (Westgate, 1968). The hummocky terrain south of the West and Center blocks contains ice-walled lake plains and numerous eskers. These former lakes and eskers emptied through short channels into the Middle Creek channel. Outwash is observed at the west arm of Cypress Lake and glaciolacustrine sediment covers the glacial deposits at Consul.

South of the Cypress Hills, the early Middle Creek channel cut during retreat of the ice from the area south of the West and Center blocks (Fig. 6.9, 6.10) southeast of

Merryflat was partially infilled (Fig. 6.9, 6.10). The ice reached the Frenchman channel and deposited a small ridge of sediment in the Frenchman channel north of West Plains (Fig. 6.10) and redirected the meltwater flowing in the Battle Creek channel eastward along the Frenchman channel to the west arm of Cypress Lake depositing outwash there (Fig. 6.10). The meltwater then flowed southwards along the margin of the ice and collected in Lake Consul (Fig. 6.10). Silt and clay rhythmites exposed in sections along the Battle Creek at Consul, Saskatchewan were deposited in this lake. Drainage south from Lake Consul cut the channel presently occupied by Battle Creek. Lake Consul did not drain through catastrophic outburst but by gradual outflow as shown by the numerous terraces, slip-off slopes, point bar, and channel deposits associated with this channel. The terminus of the West Lobe in this area is uncertain but because the Battle Creek sections show no indication of being over-ridden, it is unlikely to have reached this area. Meltwater, flowing southeastwards along the west flank of the West Block, flowed along the Medicine Lodge channel. This channel drained into Lodge Creek. The Middle Creek channel became a hanging valley above the Medicine Lodge channel and was abandoned as a meltwater conduit (Fig. 6.11).

Along Battle Creek, glaciofluvial sediment rests unconformably on an erosion surface cut into the subglacial till (Chapter 2, this volume). These terrace deposits are unfaulted and undeformed. The debris-laden ice from which the underlying subglacial meltout till was derived, had therefore completely melted before deposition of the glaciofluvial sediment. Clayton (1967), Driscoll (1980), Ostrem (1959), Sharp (1949), and Watson (1980) have observed that buried ice can persist for thousands of years. A significant time interval therefore, likely had elapsed between the Underdahl and Middle Creek advances.

North of the Cypress Hills in Saskatchewan small channels at about 1035 m elevation may mark the edge of a restricted Gap Lobe at this time (Fig. 6.11). Ice-marginal channels at an elevation of about 1065 m along the north flank of the East Block may indicate the Middle Creek terminus there. Ice-thrust bedrock masses that form the up-ice parts of the Dollard drumlins (Fig. 6.2, 6.13) may have thrust at this time or during the Underdahl Advance.

West and north of the West Block of the Cypress Hills, the limits of the Middle Creek Advance are uncertain and are shown with dashed lines (Fig. 6.4). The movements of the ice mass north of the Cypress Hills (from which the Gap Lobe and East lobes extended), are also uncertain. As Westgate (1968, 1972) did not describe this advance in his glaciation chronology, it is unclear which of his drifts, if any, correspond to the Middle Creek Advance. Deformed glaciolacustrine sediments exposed in a pit 1 km north of Elkwater Lake (Fig. 6.9) were interpreted by Vreeken (1986) as pro-glacial lake sediments pushed by ice advance at the Late Wisconsinan maximum. It is possible that the sediments were deposited in a lake formed during retreat of the Underdahl ice and deformed during the Middle Creek readvance. The Middle Creek Advance in the Elkwater area would thus closely match the limits of the Underdahl Advance. Glacier Peak ash of either the Manyberries event (12750 ± 350 BP, Westgate and Evans 1968) or the Chiwawa event (11200 ± 100 BP, Westgate and Briggs 1980) was found in closed depressions in the hummocky terrain zone (Vreeken 1986). The hummocky terrain is therefore older than these ash layers but exactly how old is uncertain.

If Vreeken's (1986) conclusions are correct then either the limit of the Middle Creek ice is farther north and has yet to be found, or the ice north of the Cypress Hills did not retreat from the Underdahl limit like the West Lobe did. If the ice north of the Cypress Hills therefore did not retreat from the Underdahl limit it is likely that it also did not undergo the Phase D retreat and Phase E advance described in Clayton and Moran (1982). Identifying the correct scenario will only be possible when the relationship between the West Lobe and the ice north of the Cypress Hills is more fully understood and the position of the suture zone between the ice north of the Cypress Hills and West Lobe has been identified. If the two ice masses did not respond synchronously to external forces then the possibility of miscorrelating events south and north of the Cypress Hills is very strong.

The revised glaciation sequence presented here differs from that of Westgate (1968) and Christiansen (1979) who indicate that once the ice had receded from its Late Wisconsinan maximum position, it did not re-enter the area south of the Saskatchewan Cypress Hills. The Middle Creek Advance margin south of the Cypress Hills, as presented here, is less extensive than the maximal margin of Phase E (about 16,600 BP) of Clayton and Moran (1982). As discussed the existence of a Phase E advance north of the Cypress Hills, following a retreat of unknown extent, needs further examination. This point will be revisited in the discussion.

Summary

During Time III the West Lobe readvanced to the Middle Creek channel on the south margin of the West Block (Fig. 6.10). This ice over-rode the Middle Creek channel formed during the Time IIb retreat, partially infilling it. The advance also deposited a small ridge of sediment in the Frenchman channel northeast of West Plains (Fig. 6.10). The West Lobe terminated at the west end of Cypress Lake. Meltwater flowed into the Frenchman channel, depositing outwash at the west arm of Cypress Lake. Meltwater following the margins of the advance was impounded in Lake Consul.

Southward drainage from this lake formed the channel occupied by the modern Battle Creek, depositing glaciofluvial sediment in high-level terraces. The location of the southern terminus of West Lobe is uncertain, but the ice did not override the Battle or Lyons Creek sections near the Montana border.

The northern ice mass and the East Lobe again appear to have been relatively immobile. Ice thrusting of glaciolacustrine sediment at Elkwater Lake, Alberta, and bedrock on the Swift Current plateau may have occurred at this time.

Retreat from the Middle Creek Advance terminus is poorly understood. Meltwater, from the wastage of the West Lobe, flowed along the Medicine Lodge channel (Fig. 6.11). The extent and duration of the retreat are unknown.

Time IV. The Altawan Advance. About 15,000 to 14,000 BP

The Altawan Advance, is observed along the border of Alberta and Saskatchewan (Fig. 6.3, 6.11), where it partially infilled the Medicine Lodge channel. The advance terminated at about 915 m elevation and extended only a few kilometers into Saskatchewan. The limit of this advance is marked by a zone of hummocky and ridged terrain that deflects Lodge Creek eastward near the Montana border. Near Altawan, the surface has been scoured to bedrock, leaving erratics on the shale surface. Glacial sediment at the Altawan Reservoir consisting of layers of silt and fine sand interbedded with diamicton beds with few stones. These layers are commonly contorted and some of the beds have vertical to near-vertical orientations. These contortions indicate substantial reorientation, following the meltout of buried ice blocks. If the ice blocks were remnants from the Middle Creek Advance buried beneath sediments deposited during the Altawan Advance, then only a short time interval (a few hundred years?) would have passed between the Middle Creek and Altawan advances.

The Altawan Advance was not recognized by Westgate (1968) but he noted that the ice had undergone several minor readvances during retreat from his Wildhorse (Green Lake) margin before it reached the Pakowki margin but supplied no further information. The Altawan Advance may have been one of these small readvances during extended retreat from the Middle Creek margin. The infilling of the Medicine Lodge channel indicates that the Altawan was a distinct advance. Its recognition does not require major alterations to the existing deglaciation sequences.

During retreat from this margin, meltwater flowed southeast along Jaydot Coulee and then Lodge Creek. This enhanced the scabland appearance of the terrain in the extreme southwest corner of Saskatchewan. The Milk River channel (Fig. 6.11) was also cut during the retreat from this position and Lake Wildhorse was impounded in the extreme southeast corner of Alberta (Fig. 6.11). This interpretation differs from Westgate's interpretation (1965, 1968) in the lake was impounded behind eskers formed during retreat from the maximum limit of the Late Wisconsin glaciation. Because the area was covered by the Middle Creek Advance, the lake could not have formed before this time. By the end of the Altawan Advance, the ice had retreated permanently from the area south of the Saskatchewan Cypress Hills.

No field evidence showing movement of the main ice mass north of the Cypress Hills or the Eastern Ice Lobe has been found. The location of the ice front, north and east of the Cypress Hills, is unclear (Fig. 6.3, 6.11).

Clayton and Moran's (1982) chronology does not identify an advance at this time but their chronology's broad regional scope cannot incorporate minor local readvances. Their Phase F to H margin (about 14,000 to 13,500 BP), which they tentatively correlated to the subsequent Pakowki moraine, is an amalgamation of three phases. It is possible that one of the earlier phases (Phase F or G) correlates to the Altawan Advance. Phase H would then correlate to the Pakowki moraine.

Summary

The Altawan Advance partially infilled the Medicine Lodge channel north and west of Altawan and extended only a few kilometers into southwest Saskatchewan (Fig. 6.11). Meltwaters from this advance deposited outwash and eroded the broad scabland immediately east of the narrow belt of hummocky terrain along the border of Alberta and Saskatchewan. The advance deposited diamicton and glaciofluvial sediment over ice at the Altawan Dam and near Govenlock. During retreat from the Altawan margin, Lake Wildhorse was impounded, the Milk River channel formed, and meltwater flowed along Jaydot and Lodge creeks.

Time V. Pakowki Advance. About 14,000 to 13,500 BP.

In the Alberta part of the study area, the Pakowki Advance margin follows the Pakowki margin of Westgate (1968). There, the limit of the Pakowki Advance ice is located between the Etzikom Coulee and the Milk River channel at a prominent ridged end moraine along the north side of the preglacial Comrey bedrock high (Westgate 1968, Barendregt 1977), (Fig. 6.3, 6.12). The ice reached an elevation of about 975 m and covered much of the dissected terrain on the west side of the West Block. The discontinuous ridge, north of the West Block, that marks the Pakowki limit (Westgate 1968), was traced by the author, eastward into Saskatchewan, to near the east flank of the

East Block (Fig. 6.3, 6.12) . On the Swift Current plateau in the area affected by the East Lobe, no ridges or landforms could be directly linked to this ridge system. Jones Creek, (an ice-marginal channel) or the Eastend Coulee (a spillway channel), (Fig. 6.12), may mark the limit of the advance in the Swift Current plateau area, but this is uncertain. Westgate (1968) identified several channels and coulees that were partially infilled during the Pakowki Advance. The Pakowki therefore was a readvance and not a stillstand.

During retreat, Lake Pakowki (Fig. 6.13) was dammed between the bedrock high and the active ice mass (Westgate 1968; Barendregt 1977). Barendregt (1977) noted that the lake drained over the divide first through Canal Coulee, then Wild Horse Coulee (Fig. 6.12, 16.13), and finally along the Pakowki channel into the Milk River channel. In Barendregt's sequence, the lake was considered to be a pre-Late Wisconsinan feature. Whereas in Westgate's and the present author's sequences, the lake is considered to be Late Wisconsinan.

Sometime after the ice retreated, the Manyberries Ash was deposited over the Pakowki drift (Westgate 1968). The ash came from Glacier Peak and has been dated about 12,500 BP (Westgate and Evans 1978). More recent analysis of the ash have arrived at a 11,200 BP date (Mehringer et al. 1984). Deformation of the ash at its type locality indicates that it was deposited over sediment containing buried ice blocks (Westgate 1968). Westgate (1968) estimated that the Pakowki sediment could therefore be as much as 2000 years older than the ash itself. Westgate and Evans (1978) also identified Glacier Peak Ash near Irvine, Alberta. Using an average rate of sedimentation from a dated horizon they determined that the Irvine Ash was 13,750 BP. As mentioned Vreeken (1986) identified Glacier Peak Ash in enclosed depressions near Elkwater. These ash deposits provide the only definite Late Wisconsinan timeline in the study area.

A large lake, impounded north of the Cypress Hills during retreat (Fig. 6.13). was called glacial Lake Downie (Vreeken 1989). He estimated the age of its drainage as 13,510 BP. Initially it drained along Jones Creek and then along Eastend Coulee into the Frenchman channel.

Vreeken (1991) identified subglacial drainage features on the Swift Current Plateau. The Dollard drumlins (Fig. 6.2, 6.13) on the plateau contain ice-thrust bedrock with glaciofluvial sand and gravel over them and on their downice sides. What appear to be fluvial scour marks are also observable around the drumlins. It is possible that subglacial drainage of the lake impounded north of the Cypress Hills deposited sand and gravel in the lee of these pre-existing thrust blocks (thrust during the Underdahl or Middle Creek advances?) and eroded scours around them. The lake may have undergone several of these drainage events. The local jokulhlaup outburst hypothesis is simple and explains all the features. Erosion of subglacial landforms and features formed by the East Lobe, by the outburst(s), may have removed the evidence needed to reconstruct the glacial sequence in the area. An alternative hypothesis Rains et al. (1993) and Shaw (pers. com. 1994) is that a non-local mega-flood originating in northern Alberta formed the drumlins and eroded the scours around them.

Christiansen (1979) and Dyke and Prest (1987b) did not include this stage in their deglaciation sequences. Barendregt (1977) believed the Pakowki moraine was deposited by an Illinoian or older ice sheet. As previously mentioned, this view has been challenged by Clayton and Moran (1982), Shetsen (1984), Fenton (1984), and Fullerton and Colton (1986). Clayton and Moran (1982) tentatively equated the Pakowki moraine to their Stages F to H, about 14,000 to 13,500 BP.

Summary

The Pakowki Advance did not reach Montana (Fig. 6.12) but terminated west and north of the Cypress Hills. The ice reached the flanks of the preglacial bedrock divide (Fig. 6.2) forming a ridged moraine. North of the Cypress Hills the margin of the advance may be along a discontinuous ridge system (Fig. 6.12). No evidence for movement of the East Lobe has been found. During retreat from this margin, meltwater was impounded between the bedrock high and the ice, forming Lake Pakowki (Westgate 1968) The lake drained first along Canal Coulee, then Wildhorse Coulee, and finally through the Pakowki channel. North of the Saskatchewan Cypress Hills, another large lake (Lake Downie) formed during the retreat. This lake may have undergone periodic jokulhlaup drainage events that contributed to the subglacial formation of the Dollard drumlin field. Late drainage flowed along Jones Creek and when the ice receded further, a lower outlet, the Eastend Coulee, was used.

Time VI. The Etzikom Advance. About 12,300 BP

The Etzikom Advance (Fig. 6.13) influenced the extreme north edge of the study area. During this advance, the Granlea and Etzikom channels in Alberta were partially infilled, and the Etzikom drift was deposited (Westgate 1968). The infilling of these channels indicates that the Etzikom Advance was an advance and not as stillstand. In Alberta, Westgate (1968) placed the Etzikom Advance margin along a broad, well-developed end moraine at about 910 m elevation (Fig. 6.3, 6.13). The well-developed hummocky terrain (Klassen 1991), northwest of Maple Creek, along Highway 1, was produced by this advance.

Drainage from this margin was to the northeast along Norton channel and Seven Person's Coulee (Fig. 6.13). Lake Medicine Hat (Fig. 6.13) formed during retreat from the Etzikom margin. The advance did not reach or overtop the Comrey preglacial bedrock high (Fig. 6.2, 6.12).

A large delta formed northeast of Maple Creek, Saskatchewan, (Fig. 6.13) and prograded into remnants of Lake Downie or a lake formed during the early stages of retreat from the Etzikom margin. The delta contains gravel and coarse sand in the northwest and fines to the southeast. Five to eight metres of silt and clay rhythmites are observed in sections along Gap Creek. South and east of Maple Creek glaciolacustrine sediments cover much of the surface (Klassen 1991). Hummocky lacustrine sediments (Klassen 1991, 1992) in the former lake basin indicate that the sediment in this lake in places covered stagnant blocks of ice. Initially the lake drained southwards along Jones Creek and then through Eastend Coulee into the Frenchman channel, but as the ice wasted back, the lake drained eastward along Swift Current Creek into Pelletier channel north of the Wood Mountain Upland (Fig.6.2, 6.14F).

Stalker (1977), Stalker and Harrison (1977) and Barendregt (1977) linked the Etzikom moraine to the Lethbridge moraine, and concluded that the Etzikom moraine was the maximum limit of the Late Wisconsinan ice in the western Prairies. In contrast, Westgate (1968, 1972), Christiansen (1979), Clayton and Moran (1982), Fullerton and Colton (1986), Vreeken (1986, 1989), and the present author considered it to be a retreatal position of the Late Wisconsinan ice. Vreeken (1989) dated the Lethbridge moraine as 13,510 BP.

Summary

The Etzikom advance affected only the northwest part of the study area. It infilled small channels and impounded a large lake north of the Cypress Hills. Sediment deposited in this lake covered much of the terrain north of the Cypress Hills. Initially the lake drained southeastwards via small creeks into the Eastend Coulee and then into the Mississippi drainage system. Subsequent wastage of the ice, allowed the lake to drain northeastwards along Swift Current Creek and then into the Pelletier Channel.

Discussion

Timing of the deglaciation sequence

Clayton and Moran's (1982) Late Wisconsinan glaciation chronology is the most reliably dated because only C14 wood dates were used to limit contamination problems.

(Non-wood dates, i.e., from gytta and disseminated organic matter, are often anomalously old due to incorporation of crushed lignite and shale, sources of old inert carbon, that are disseminated throughout the sediment in the region; Clayton and Moran 1982; MacDonald et al. 1986). The scarcity of C14 wood dates from the western Canadian Prairies and Montana hinders the extension of their chronology into the study area. The validity of Clayton and Moran's (1982) chronology therefore depends on the reliability of their long-range correlations but because the movement of ice sheet margins was not synchronous along their entire lengths, a large element of uncertainty enters into the correlations. The correlation of boundary positions is also hindered by the non-uniform manner in which sediment is eroded and deposited by an ice sheet. This leads to numerous local depositional and erosional gaps that complicate the reconstruction of events.

In Clayton and Moran's (1982) chronology, the onset of glaciation in the Western Plains occurred between 28,000 and 24,000 BP. Phase D (Time I this study) timing is not well established but their regional inferences provide an estimate of between 18,000 to 20,000 BP. Their Phase E (Time III, Middle Creek Advance?) also lacks local dates, and while their regional correlations yield a 15,000 BP age for this phase, it may be several thousand years older (Clayton and Moran 1982). Their Phases F through H, between 14,000 and 12,000 BP, were correlated to the Robinson (Pakowki) Advance (Time V). The Etzikom Moraine (Time VI) was equated to Phase I (Clayton and Moran 1982) occurring between 12,300 to 11,700 BP.

Clayton and Moran's (1982) chronology is several thousand years longer than that proposed by Christiansen (1979) but several key dates used by Christiansen (1979) were from gytta and disseminated organic material and may have been contaminated. Teller (1989) noted that the dates for the Rossendale Site, that were largely responsible for the rapid deglaciation chronology, have proven to be unreliable. New dates of 9600±70 BP and 9510±90 BP from wood (Teller 1989), support the slow retreat chronology advanced by Clayton and Moran (1982) and Moran and Clayton (1983).

Dyke and Prest (1987b) placed their Late Wisconsinan maximum at 18,000 BP which is at the low end of the time frame used by Clayton and Moran (1982). By stage two of Dyke and Prest (1987b), about 14,000 BP, the ice had already retreated to the Lethbridge moraine (Time VI) west of the Cypress Hills and the Etzikom margin (Time VI) north of the Cypress Hills and Wood Mountain Upland. Neither Christiansen (1979), nor Dyke and Prest (1987a) discussed local readvances.

Wood dates of between 12,000 to 6,000 BP, obtained by Christiansen and Sauer (1988) from colluvium near the base of the Frenchman channel led them to conclude that the Frenchman channel was a Late Wisconsinan feature. Dates obtained by Klassen and Vreeken (1987) from surface materials surrounding the Saskatchewan Cypress Hills, and tabulated in Klassen (1992) are all younger than 13,000 BP even though many are from disseminated organic matter and apt to give anomalous old dates. This indirectly supports a Late Wisconsinan age for the uppermost glacial sediments south of the Pakowki moraine.

Glacier Peak ash is found in several locations in the study area. The Manyberries Ash (about 12,500 BP, Westgate 1968; 11,200 BP, Mehringer et al. 1984) was deposited in sediment covering ice derived from the Pakowki Advance. The Irvine Ash (13,750 BP, Westgate and Evans 1978) is apparently older than the Manyberries Ash to the south. This discrepancy may be due to errors in the assumed average sedimentation rate that was used to place a date on the ash layer (Westgate and Evans 1978). The Glacier Peak Ash described by Vreeken (1986) may be equivalent to either the Manyberries event (12,750 ±350 BP) or the Chiwawa event (11,200± 100 BP). The ashes provide minimum ages for the surfaces but the exact ages of the surfaces are still unknown.

Late Wisconsinan ice lobe distribution and movements

During Time 1 of Christiansen (1979) and Phase D of Clayton and Moran (1982) the East Lobe was placed north of the Frenchman channel while the West Lobe extended to the south bank of the channel (Fig. 6.5). Meltwaters draining from these lobes gradually incised an interlobate Frenchman channel. The Elkwater drift (Fig. 6.3), (Westgate 1968, 1972) and the "bedrock terrain with residual drift landscape complex" (Fig. 6.1), (Klassen 1992) therefore lay beyond the limits of the last ice, and could be interpreted as the remnants of an older, more extensive till sheet.

In the glacial sequence advanced in this paper, the Laurentide Ice Sheet reached its maximum during the Underdahl Advance. During that time, the Gap Lobe extended across the Center Block of the Cypress Hills and coalesced with the East and West lobes near Cypress Lake (Fig. 6.7), (Fig. 6.8). In the new ice lobe configuration, the East Lobe extended westward over the Frenchman channel area. The Gap Lobe is restricted to the south edge of the Center Block and does not extend southward into Montana as shown in Dyke and Prest (1987a). The West Lobe was confined to the plain west of the Old Man On His Back and Boundary plateaus (Fig. 6.7). The three lobes coalesced in the Cypress Lake area. Glaciolacustrine sediment and several short spillways on the southern flanks of the Saskatchewan Cypress Hills indicate that, instead of small ponds as suggested by Vreeken (1986), several large pro-glacial lakes were impounded on the southern flanks of the Cypress Hills (Chapter 3, 4, this volume). Ice rafting across these lakes distributed erratics beyond the limit of the Underdahl Advance. The Elkwater drift (Fig. 6.3) and "bedrock terrain with residual drift" (Fig. 6.1) are therefore Late Wisconsin ice-raft deposits. Westgate (1968) questioned whether the Elkwater drift was truly an older deposit or one related to the Late Wisconsin glaciation. He subsequently correlated the Elkwater drift with the highest drift surrounding the Del Bonita Upland solely on their similarity in elevation. Because the highest drift around the Del Bonita Upland was interpreted to be pre-Late Wisconsinan (Stalker 1962), the Elkwater drift also became pre-Late Wisconsinan. The new interpretation shows that the Elkwater drift and Underdahl Advance are inter-related.

In the new configuration, the Frenchman channel area was covered by ice at the maximum of the Late Wisconsin glaciation. The channel therefore, was not an interlobate feature gradually eroded by meltwaters draining between the ice lobes as shown by Christiansen (1979) and Clayton and Moran (1982), (Fig. 6.5). The Frenchman channel is reinterpreted as a spillway rapidly incised by the catastrophic outburst of the lakes impounded on and around the southern flanks of the Cypress Hills (Fig. 6.8) (Chapter 4, this volume). The sidehill position of the channel and the absence of a southern wall in some locations indicates that ice formed the south wall of the channel during its incision. The new distribution pattern requires that ice cover the "bedrock terrain with drift landscape complex" (Fig. 6.1), (Klassen 1992) and that ice from the northeast, cover the "first advance drift" (Fig. 6.1), (Klassen 1992). However, the ice distribution pattern broadly agrees with Fullerton and Colton (1986) and Christiansen and Sauer (1988).

The glacial sequence of this paper recognizes two new advances, the Middle Creek and the Altawan. The limit of the Middle Creek Advance in Alberta (Fig. 6.10) and north of the Cypress Hills is uncertain. If the Middle Creek Advance corresponds to Phase E of Clayton and Moran (1982), which reached nearly the same limits as their earlier more extensive Phase D event, then the Middle Creek limit would be at or very near to the Underdahl limit. A Middle Creek advance does not require a major revision of the broad frameworks established by Westgate (1968) or Clayton and Moran (1982). South of the Cypress Hills the Middle Creek Advance is less extensive than Clayton and Moran's (1982) Phase E limit but this is a local reconfiguration.

The Altawan Advance was previously unrecognized. Its recognition does not require major alterations to the existing deglaciation sequences. Similarly the Pakowki and Etzikom advances only extend the margins of Westgate (1968) into Saskatchewan. They therefore do not require changes to the earlier chronologies.

Recognition of the Middle Creek and Altawan advances changes the earlier chronologies such as Christiansen (1979), Clayton and Moran (1982) and Vreeken (1986) reconstructions but the occurrence of these advances north of the Cypress Hills is still uncertain. The lack of information on movements of the main ice mass north of the Cypress Hills (which contains the East Lobe) during deglaciation hinders accurate reconstructions. The fragmentary deglaciation picture available for the area north and east of the Cypress Hills may reflect the compression of a number of events into the 30 km zone between the limits of the Underdahl and Etzikom advances. Oscillations of the ice sheet in such a confined area may have removed much of the evidence regarding the advances and retreats. Without absolute dates the correlation of ice marginal positions observed there is reduced to educated finger-matching. A large number of speculative correlations is therefore possible.

The other possibility that the ice of the West Lobe was not coupled to the ice north and east of the Cypress Hills and behaved independently, has already been mentioned. The West Lobe of this paper corresponds to Shetsen's (1984) Center Lobe (Fig. 6.4), and originated in the Keewatin area of the Northwest Territories (Fig. 6.4) as indicated by the oolitic ironstone erratics from the east end of Great Slave Lake (Klassen 1989) that are present only in areas formerly covered by the West Lobe . The main ice mass north of the Cypress Hills corresponds to the East Lobe of Shetsen (1984) and most probably originated in central or Northern Manitoba. In contrast to the numerous ice-marginal positions observed in the area affected by the West Lobe, the main ice mass north of the Cypress Hills appears to have been essentially immobile. Perhaps it was blocked by the mass of the Cypress Hills or slowed as it flowed onto the Missouri Coati. This more stable ice formed fewer ice-related landforms. Its long stillstand on the north side of the Wood Mountain Upland cut the extensive meltwater channel system observed there.

In contrast to the stable northern ice mass, the numerous ice-marginal positions occupied by the West Lobe may indicate that it was a surging ice lobe similar to those described in other areas along the Late Wisconsinan margin (Clayton et al. 1985). Evans and Campbell (1992, p. 550) report that the upper part of their LFA5 unit in Dinosaur Provincial Park near Brooks, Alberta, has features that suggest that the ice had become decoupled from its bed and had probably surged. Surging of the West Lobe could have formed the flutes and drumlinoid ridges west of the Cypress Hills (Westgate 1968; Barendregt 1977) and the large flutes near Brooks, Alberta (Campbell 1993). A decoupled, surging West Lobe could also explain the rapid drop in elevation of erratics, from 1650 m on the north side of the Sweet Grass Hills (Barendregt 1977) to 1450 m on the north side of the West Block (Westgate 1968). The West Lobe and the ice mass north of the Cypress Hills may have been fundamentally different. Further research to test this conclusion is needed.

The glaciation sequence presented here, depicts the ice of the Late Wisconsin ice as the most extensive to affect the study area. This agrees with studies from central Alberta (Liverman et al. 1989; Young et al. 1989, 1993, 1994; Burns et al. 1993) that obtained mid-Wisconsinan dates from wood and bone in preglacial sediment in the Peace River and Edmonton areas respectively. More extensive Late Wisconsinan ice means that the ice-free corridor was more restricted than envisioned by Stalker (1980). That the Late Wisconsinan was the most extensive ice in the area does not agree with the observations of Fullerton and Colton (1986) and Fullerton (pers. com., 1991) who maintained that several pre-Late Wisconsinan till units are present in sections in Montana. Nor does it agree with Klassen (1992), who stated that several of the surface glacial units were deposited before the Late Wisconsin Glaciation.

Fulton et al. (1984) and Evans and Campbell (1992) reported only a single pre-Late Wisconsinan event (Dunmore Glaciation) in exposures at Medicine Hat and Dinosaur Provincial Park (near Brooks, Alberta), respectively. Fulton et al. (1984) and Evans and Campbell (1992) also observed deposits from the Lostwood Glaciation (Late Wisconsinan) at Medicine Hat and Dinosaur Provincial Park respectively. Glacial sediment, deposited during the Underdahl Advance at the Battle and Lyons creeks, is the temporal equivalent of the Late Wisconsinan Loring and Fort Assiniboine tills that cover the northern Montana plains (Fullerton and Colton 1986). The regional distribution of the Late Wisconsinan ice (Fig. 6.14) indicates that the most likely area to find pre-Late Wisconsinan material at the surface is on the south side of the Wood Mountain Upland northeast of the Frenchman channel (Fig. 6.14). Fieldwork along Bluff Creek (Fig. 6.14) in this area (beyond the limit of the main study area) has revealed a deeply weathered chocolate brown diamicton that may be older till (Kulig, unpublished data). The glacial stratigraphy in Alberta therefore seems to be much less complicated than that described by Fullerton and Colton (1986).

The lack of pre-Late Wisconsinan glacial deposits on and around the Cypress Hills is problematic. These deposits may have been completely eroded, not recognized, or the flow patterns of earlier ice may not have matched those of the last ice sheet. The absence of earlier deposits may be related to the formation of an independent West Ice Lobe during the Late Wisconsinan. If an ice divide developed farther to the west during the Late Wisconsinan than had developed during earlier glaciations ice coverage during the Late Wisconsin glaciation over the study area and throughout Alberta would have been significantly altered. This ice divide would be similar to the Caribou Hills ice divide during the Late Wisconsin glaciation (Dyke et al. 1982; renamed the Plains ice divide by Dyke and Prest 1987b). Its formation would explain the restriction of oolitic ironstone erratics from east of Great Slave Lake to the west side of the Cypress Hills as only this area was covered by the West Lobe. It would also account for the difference in chronologies between the West Lobe and the ice mass north of the Cypress Hills as a West Lobe derived from an separate ice divide could advance and retreat or surge independently from the main ice mass.

Pre-Late Wisconsinan till units are located along the South Saskatchewan River near Medicine Hat (Stalker 1976; Proudfoot 1985; Evans and Campbell 1992). Mid-Wisconsinan dates from between two till units in Evil Smelling Bluff near Medicine Hat (Stalker 1976) in southern Alberta indicate that an earlier ice mass had entered at least that far into Alberta. Evidence for an earlier ice mass in Alberta is also present in Elk Island National Park, 40 km east of Edmonton. There, Jennings (1984) obtained Mid-Wisconsinan range radiocarbon dates from between two till sheets. The total westward extension of this older ice mass into Alberta is uncertain. The ice centers from which this earlier ice had spread west were probably significantly different from those of the Late Wisconsin glaciation. There is no reason to expect them to be otherwise. A change in location of ice divides between the Late Wisconsinan and pre-Late Wisconsinan glaciations would give unique ice sheet distributions and sedimentation patterns. Similarly, there is no reason to expect the same pattern of glacial events in southeast Alberta, southwest Saskatchewan and northern Montana as observed in central Canada and the U.S. mid-west. Local chronologies should not try to slavishly reproduce glacial sequences from other areas.

The revised glaciation pattern adds details to the early stages of the Late Wisconsinan deglaciation but several difficulties exist when reconstructing the regional picture. More investigation is required to determine the number and location of ice lobes in Alberta during the Late Wisconsin glaciation. Work is also required to determine the interaction of these lobes. Also no recent summary of the Late Wisconsinan glacial history of Montana containing the retreatal positions is available. Bretz (1943) and Horberg (1952) depicted several bands of hummocky terrain that probably mark retreatal positions. Lemke et al. (1965) described two advance positions. Barendregt (1977) mentioned several discontinuous moraine ridges in northern Montana. Fullerton and Colton (1986) mentioned four Late Wisconsinan ice retreatal positions in Montana but did not illustrate or discuss them. This information is needed to complete the picture. Additionally correlation between Alberta, Saskatchewan, and Montana is hampered by the different units used to classify glacial deposits in the three areas. Fullerton and Colton

(1986) correlated their lithostratigraphic units to morphostratigraphic units used in Alberta. For example, the Elkwater drift, a morphostratigraphic unit (Westgate 1968, p. 69), was correlated to their Marples Point Till, a lithostratigraphic unit. Since these units are not equivalent, the correlation's are at best speculative. A common stratigraphic approach would go a long way to clearing up this problem.

This paper is a first attempt at bringing together a wide range of data. There are many problem areas but it is hoped that the ideas and conclusions will spark further research.

Conclusions

The glaciation sequence presented here is a substantial revision of earlier Late Wisconsinan glaciation sequences for the Cypress Hills of Alberta and Saskatchewan. The Elkwater drift (Westgate 1968) is reinterpreted to be a Late Wisconsinan ice-rafted deposit. Klassen's (1992) "residual drift" that mantles the bedrock is also more probably an ice-rafted deposit, deposited during the early stages of wastage of the Late Wisconsin ice. The occurrence of an older, more extensive ice sheet to explain these units is no longer required. The chronology also contains three newly named advances: the Underdahl, the Middle Creek, and the Altawan. The Underdahl Advance was the most extensive advance in the area. The Middle Creek Advance was at or near Westgate's (1972) Green Lake Advance margin (same position as Westgate (1968) Wildhorse margin) north of the West Block in Alberta but only reached the Middle Creek channel on the south side of the West Block. The Middle Creek Advance extended into Saskatchewan.

Drift deposited during the Underdahl Advance is broadly equivalent to the Late Wisconsinan Fort Assiniboine, Loring, and Crazy Horse tills (Fullerton and Colton 1986). No equivalents have been determined for the drift sheets deposited during the Battle Creek and the Altawan Advances.

While the revised chronology fits into the chronology of Clayton and Moran (1982), noticeable differences exist between the revised chronology and the glacial history compiled by Fullerton and Colton (1986). A detailed chronology of the Late Wisconsin Glaciation of Montana would aid greatly in refining the glacial history of the Western Plains and clarify the correlations made in this and earlier publications.

The difference in the deglaciation patterns between the West Lobe, the East Lobe, and the main ice mass north of the Cypress Hills, is interesting. The West Lobe underwent several advances and retreats that are not reflected in the northern ice mass and

East Lobe. The larger size of the northern ice mass may have slowed its response to changes or the West Lobe may have been a separate ice body, decoupled from the northern ice mass. This second possibility has important regional implications.

The Late Wisconsinan ice was the most extensive ice to affect the Cypress Hills. This increase in coverage by the ice sheet may reflect formation of a western ice divide that had not formed during earlier glaciations. A western source for the West Lobe is indicated by the western-derived erratics found only on the west side of the Cypress Hills. From this western ice divide flowed the West Lobe (the equivalent of the Central Lobe described in Shetsen, 1984), that was independent of the main ice mass to the east. The numerous ice-margin positions associated with the West Lobe indicate that this lobe was active and perhaps prone to surging. In contrast the stable East Lobe experienced fewer oscillations of its margin. The movements of ice flowing from a western ice divide during the Late Wisconsinan would not necessarily have any connection to the larger Laurentide ice mass to the east. Therefore the extent and duration of any ice-free corridor between the mountains and the confining ice may not have any relationship to the movements and glacial chronology of the main Laurentide ice sheet (East Lobe of Shetsen 1984; Fig. 6.4). If a similar western ice divide was not formed during earlier glaciations, then ice coverage of Alberta may have been less extensive than previously believed.

	Reference							
LATE WISCONSIN GLACIATION UNITS	WESTGATE (1968)*	WESTGATE (1972)*	CHRISTIANSEN (1979)***	CLAYTON AND MORAN (1982)***	FULLERTON AND COLTON (1986, Chart 1)**	CHRISTIANSEN (1992)**	KLASSEN (1992), exact equivalents uncertain	THIS PAPER***
	ETZIKOM DRIFT	ETZIKOM DRIFT	PHASE 2	PHASES F-H, exact correlations uncertain	CRAZY HORSE, LORING AND FORT ASSINIBOINE TILLS, surface tills in Montana, (no Late Wisconsinan Subdivisions)	BATTLEFORD FORMATION, contains numerous local surface till units	Last Advance Drift Interlobate Drift First Advance Drift	ETZIKOM ADVANCE
	PAKOWKI DRIFT	ROBINSON DRIFT						ROBINSON ADVANCE
				PHASE E?				ALTAWAN ADVANCE
				PHASE E?				MIDDLE CREEK ADVANCE
	WILDHORSE DRIFT	GREEN LAKE DRIFT	PHASE 1	PHASE D				UNDERDAHL ADVANCE
PRE-LATE WISCONSIN GLACIATION UNITS	ELKWATER DRIFT	ELKWATER DRIFT	NONE MENTIONED	PHASES A, B, C	HERRON PARK TILL, MARKLES POINT TILL, KISLER BUTTE TILL, (ILLINOIAN)	FLORAL FORMATION upper till Early Wisconsinan, lower till Illinoian	BEDROCK WITH DRIFT LANDSCAPE COMPLEX	NO PRE-LATE WISCONSINAN DRIFT UNITS IDENTIFIED
					2 HAVRE TILLS, 2 PERCH BAY TILLS, ARCHER TILL, (PRE-ILLINOIAN)	SUTHERLAND GROUP pre-Illinioan, Warman, Dundurn and Mennon Formation till(s)	BEDROCK WITH RESIDUAL DRIFT LANDSCAPE COMPLEX	

Table 6.1. The chart shows some of the glacial units identified in the study area and adjacent areas. * denotes morphostratigraphic units (now called allostratigraphic units), ** denotes lithostratigraphic units, *** denotes chronostratigraphic event units, Klassen's (1992) landscape complexes have no direct equivalents buy should broadly correlate to the units indicated. Christiansen (1979) and Clayton and Moran (1982) used radiocarbon dates and long distance correlation of glacial landforms and glacial margins to set up their chronology. The use of lithostratigraphic, morphostratigraphic and chronostratigraphic terms in the area has created a complex and confusing glacial chronology for the area.

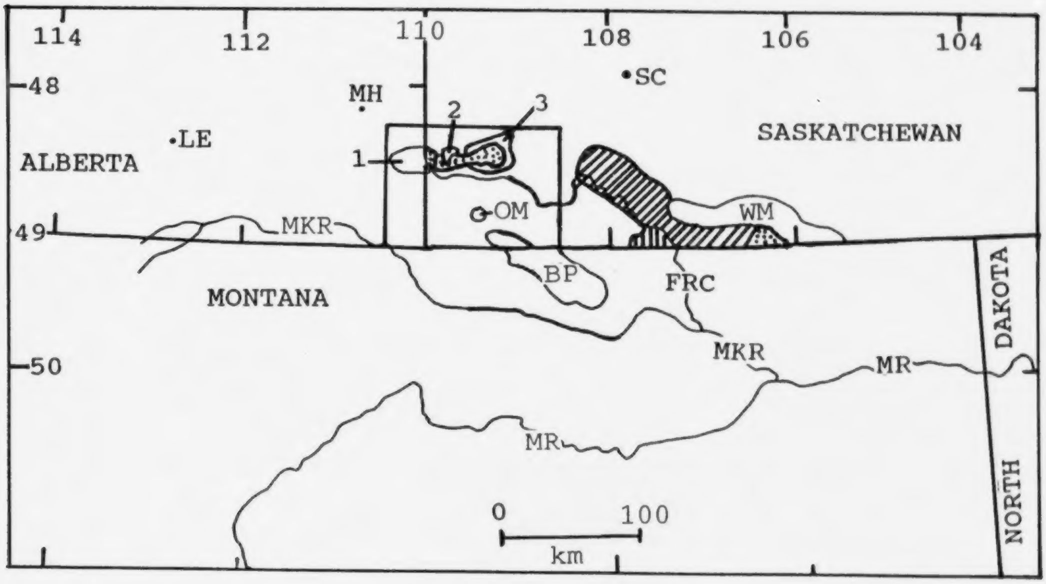
Figure 6.1. Regional setting of the study area and localities referred to in the text. The diagram also shows three landscape complexes: bedrock terrain with residual drift, bedrock terrain with drift, and first advance drift described in Klassen (1992). The bedrock terrain with residual drift was identified by Klassen (1992) as Nebraskan or Kansan, the bedrock terrain with drift as Illinoian or Early Wisconsinan, and the first advance drift as possibly the deposit of an unspecified Late Wisconsinan ice lobe from the southwest (Klassen 1992). The validity of the Nebraskan, Kansan, Illinoian, and Early Wisconsinan glaciations has been questioned (see Mickelson et al. 1986). The main study area is enclosed in the box. Letter Code: BP- Boundary Plateau, FRC- Frenchman Channel, LE- Lethbridge, MH- Medicine Hat, MKR- Milk River Channel, MR- Missouri River, OM- Old Man On His Back Plateau, SC- Swift Current, WM- Wood Mountain Upland. 1- West Block, 2- Center Block, 3- East Block.

░ bedrock terrain with residual drift landscape complex

▨ bedrock terrain with drift landscape complex

▥ first advance drift landscape complex

□ study area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MH
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LE
ALBERTA
SASKATCHEWAN
1
2
3
OM
MKR
WM
BP
FRC
MONTANA
DAKOTA
MKR
MR
MR
NORTH
0
100
km

Figure 6.2. Location map of places and features in the study area referred to in the text. Letter Code: ADC- Adams Creek Channel, AL- Altawan, ALR- Altawan Reservoir, BC - Battle Creek, BAC - Battle Creek Channel, BP- Boundary Plateau, BT- Blacktail Creek, CC- Conglomerate Creek, CL - Cypress Lake, CLX- Climax, CNC - Canal Coulee, CO - Consul, DDF- Dollard Drumlin Field, eMCC- early Middle Creek Channel, EE - Eastend, EEC - Eastend Coulee, EK - Elkwater, FR- Frontier, FRC- Frenchman Channel, FW- Fairwell Creek, GLC- Green Lake Channel, GLMH- Glacial Lake Medicine Hat, G0 - Govenlock, JAC- Jaydot Channel, JC- Jones Creek Channel, LC - Lyons Creek, LO- Lodge Creek, LTC - Lost Coulee, MB- Manyberries, MC- Middle Creek, MCC- Middle Creek Channel, MF - Merryflat, MLC - Medicine-Lodge Channel, MKR- Milk River Channel, NC- Norton Channel, OM- Old Man On His Back Plateau, PAC - Palisades Coulee, RA- Ravenscrag, RB- Robsart, SE-Senate, SH- Shaunavon, SPC- Seven Persons Coulee, SWC- Swift Current Channel, UC- Underdahl Channel, WHC - Wildhorse Coulee, WP - West Plains.

»»»»»	eskers
═	meltwater channel
⊏⊐ ⊏⊐ ⊏⊐	buried meltwater channel
←	drumlins
xxxxxxx	hummocky terminal moraine
·······	erratics limit
▨	Merryflat terminal moraine
∽ ∽	hummocky ridge deposits

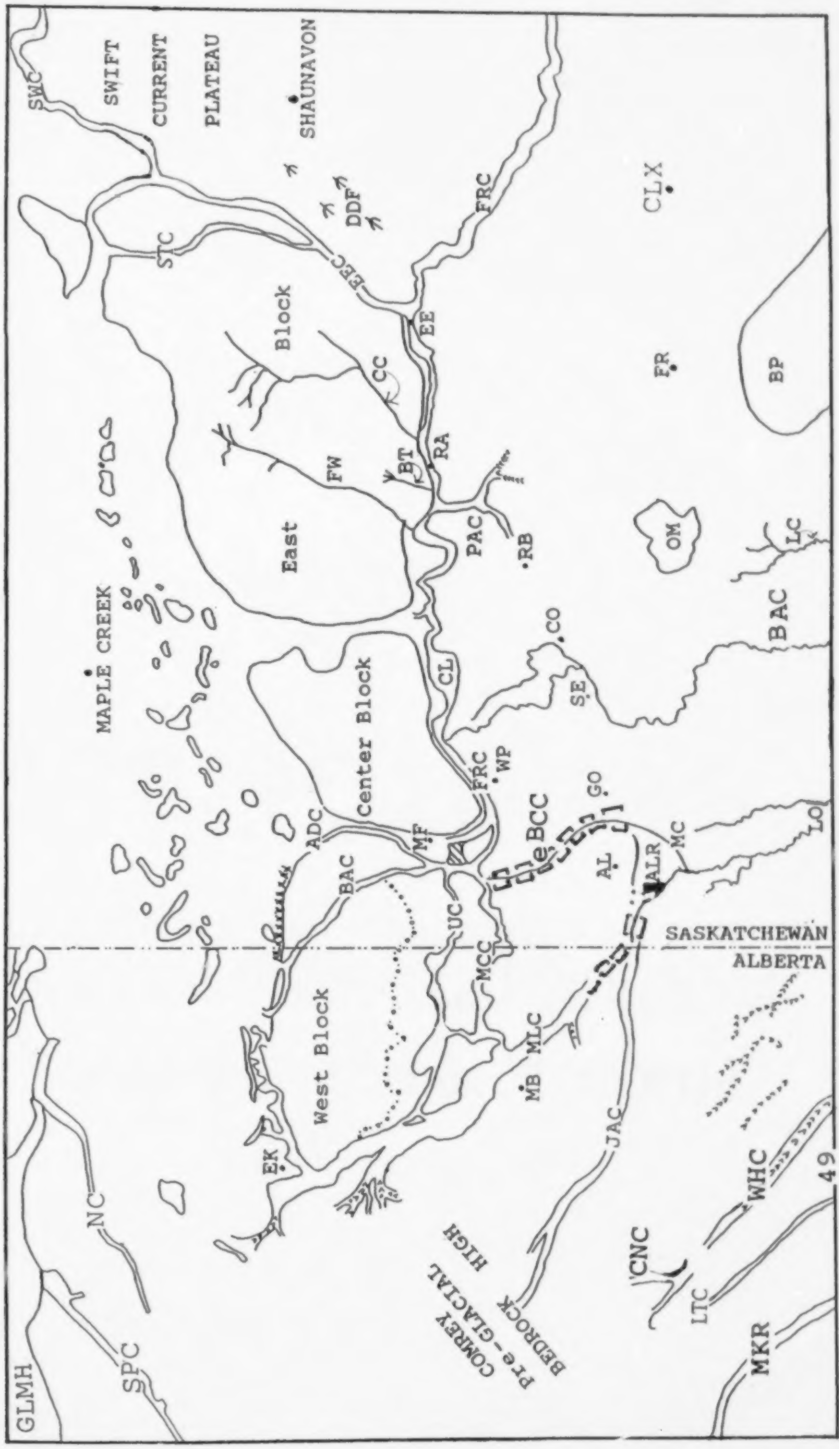
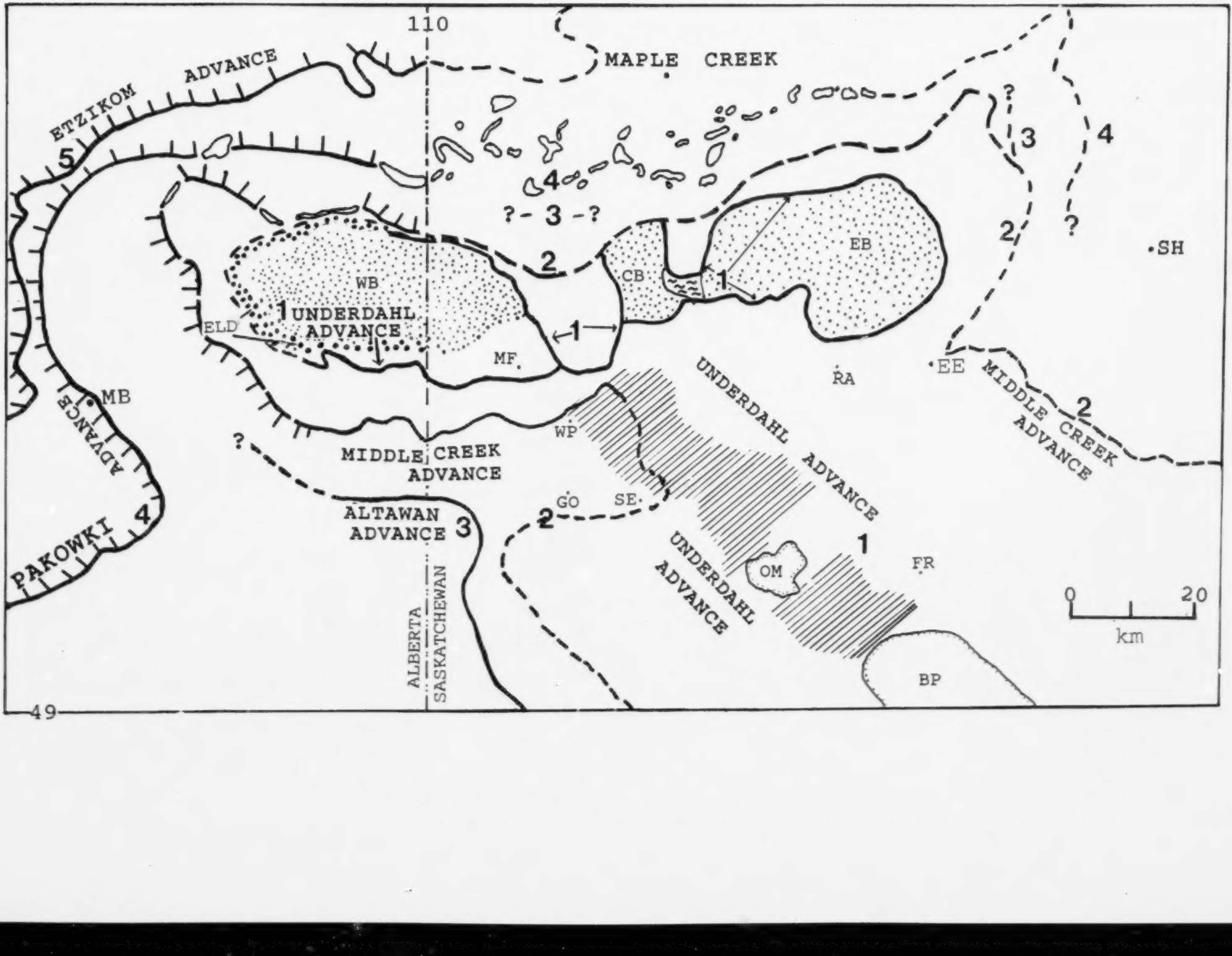

SWC
SWIFT CURRENT PLATEAU
SHAUNAVON
STC
EFC
DDF
FRC
CLX
East Block
CC
EE
FR
BP
FW
BT
RA
PAC
RB
OM
LC
BAC
MAPLE CREEK
CO
CL
SE
Center Block
FRC
WP
GO
BCC
ADC
MF
AL
ALR
MC
LO
BAC
UC
SASKATCHEWAN
ALBERTA
MCC
West Block
MB
MLC
JAC
WHC
49
EK
NC
CNC
LTC
COMREY Pre-GLACIAL BEDROCK HIGH
MKR
SPC
GLMH

Figure 6.3. The five ice advances in the study area, 1 the Underdahl Advance, 2, the Middle Creek Advance, 3, the Altawan Advance, 4, the Pakowki Advance, 5, the Etzikom Advance. Stalker's (1977) Late Wisconsinan margin followed the Etzikom Advance Margin. In Alberta the Pakowki and Etzikom margins reflect the margins drawn in Westgate (1968). The Middle Creek margin in Alberta is equivalent to the Wild Horse margin north of the West Block (Westgate 1968) but south of the West Block the Middle Creek Advance is less extensive than Westgate's Wild Horse Advance (Westgate, 1968). The location of the Middle Creek margin on west flank of the West Block is uncertain as this area may have been part of the coalescence zone between the West Lobe and the East Lobe (see text). The Underdahl Advance is the most extensive advance in the area. Letter code given in Fig. 6.2.

////	coalescence zone between the East and West Lobes
░	area not glaciated during the Late Wisconsin glaciation
┬┬┬┬	Westgate's (1968) margins
TT TT	Westgate (1968) margin uncertain
~	margins developed in this paper
- ~ -	uncertain margins (this paper)
∽ ∽	discontinuous hummocky ridge
≈≈≈	impounded glacial lake
⬭	plateau ice-covered during the maximum and then ice free
•-•-•-	erratics limit
∴∴∴	ELKWATER DRIFT ELD


110
MAPLE CREEK
ETZIKOM ADVANCE
5
4
? - 3 - ?
2
3
4
?
2
SH
EB
CB
1
WB
1 UNDERDAHL ADVANCE
ELD
1
MF
MB
ADVANCE
PAKOWKI 4
?
MIDDLE CREEK ADVANCE
ALTAWAN ADVANCE 3
ALBERTA
SASKATCHEWAN
WP
GO
SE
2
UNDERDAHL ADVANCE
UNDERDAHL ADVANCE
1
OM
RA
EE
MIDDLE CREEK ADVANCE
2
FR
BP
0 20
km
49

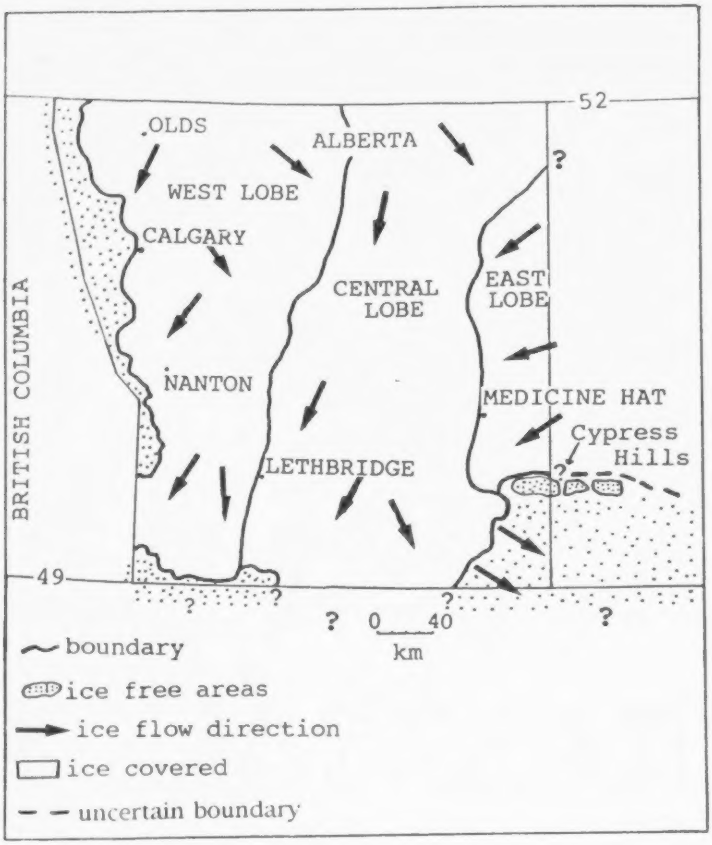
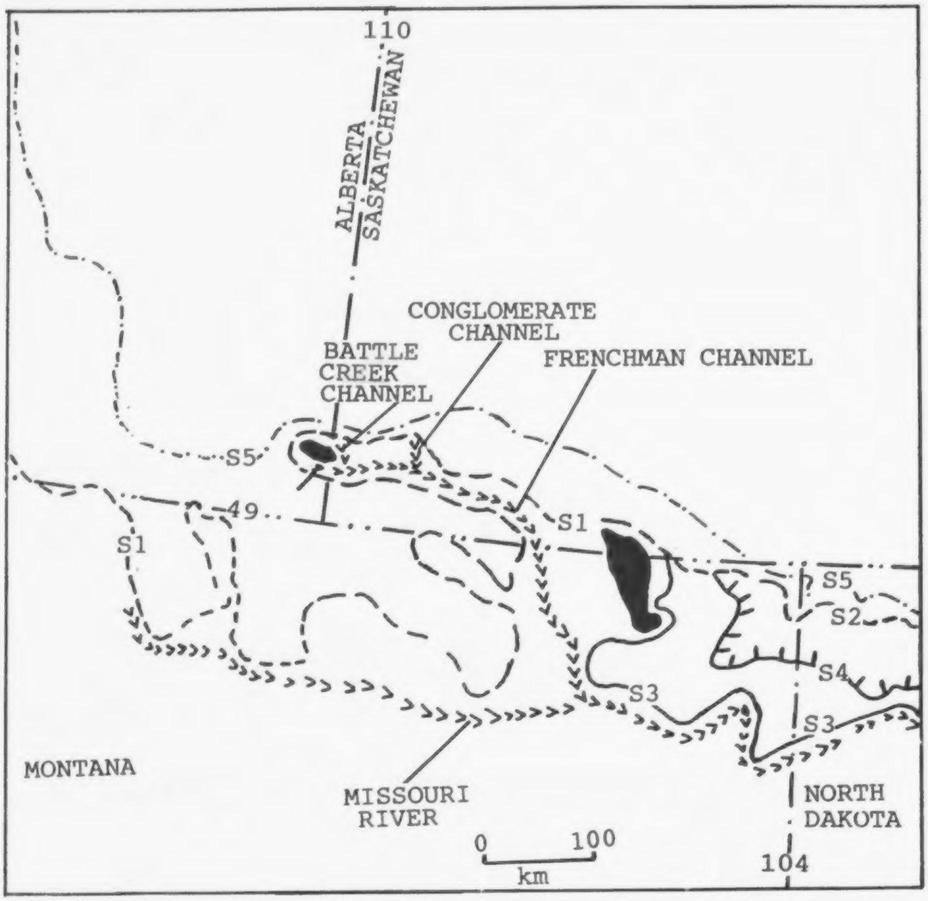

52
OLDS
ALBERTA
?
WEST LOBE
CALGARY
BRITISH COLUMBIA
CENTRAL LOBE
EAST LOBE
NANTON
MEDICINE HAT
Cypress Hills
LETHBRIDGE
?
49
?
?
?
?
?
0 40
km
boundary
ice free areas
ice flow direction
ice covered
uncertain boundary

Figure 6.4. The three ice lobes present in Alberta during the early stages of retreat of the Late Wisconsinan ice in Alberta and Saskatchewan. The boundaries of the East Lobe are uncertain in Saskatchewan as are the valleys and plateaus in British Columbia that are ice-free (after Shetsen 1984).

Figure 6.5. The distribution of ice during the maximum of the Late Wisconsin Glaciation (Christiansen's Time 1 margin) and Christiansen's Time 2 retreatal position. Time 1 of Christiansen (1979) placed the ice at the S1 and S2 limits. For their Late Wisconsinan Maximum Limit, Clayton and Moran (1982) used Christiansen's S1 limit in Alberta and Saskatchewan and northwest Montana but placed the ice at the S3 limit in northeast Montana. Clayton and Moran's Phase E readvance also attained the S1 limits in northwest Montana but in northeast Montana the ice reached only the S4 limit. Note that in both Christiansen (1979) and Clayton and Moran (1982) the ice is divided by an interlobate Frenchman Channel and that the West Lobe extends much further west than proposed in this paper. Margin S5 marks the ice margin's position at Time 2 (equivalent to the Etzikom Moraine Limit of this paper and Westgate 1968)of Christiansen's deglaciation sequence (1979).

▬	never glaciated
- - - - -	Christiansen's Time 1 margin
·—·—·—·	Christiansen's Time 2 retreat margin
∼∼∼∼	Clayton and Moran's Phase D margin
┴┴┴┴	Clayton and Moran's Phase E margin
>>>>>>>	rivers and channels


110
ALBERTA
SASKATCHEWAN
CONGLOMERATE CHANNEL
BATTLE CREEK CHANNEL
FRENCHMAN CHANNEL
S5
49
S1
S1
S5
S2
S4
S3
S3
MONTANA
MISSOURI RIVER
NORTH DAKOTA
0
100
km
104

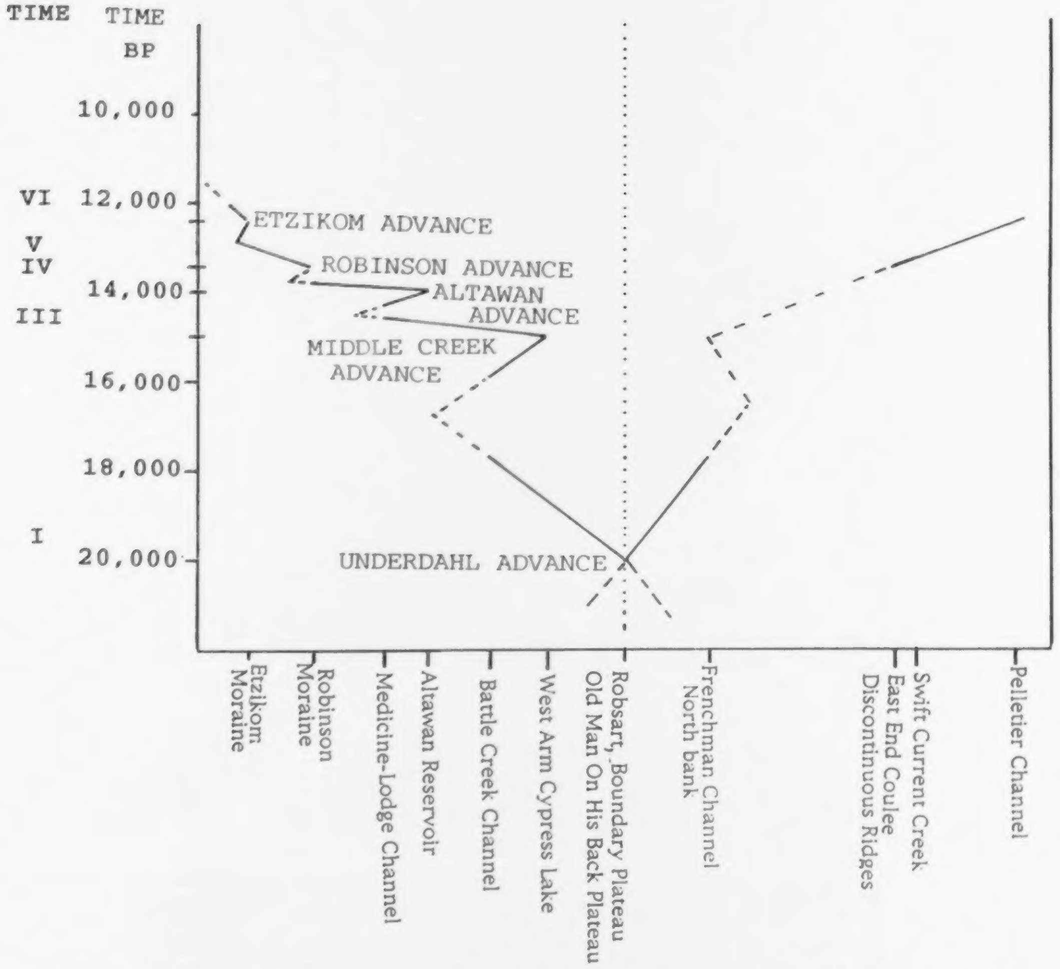


Figure 6.6. Schematic time-distance diagram of the relationship between the East and West lobes in the study area during the Late Wisconsinan. The Gap Lobe was only present during the Late Wisconsinan maximum when it coalesced with the East and West lobes. After this it has not been recognized with certainty.

Figure 6.7. TI, the Underdahl Advance, about 20,000 BP. Distribution of the ice at the maximum of the Late Wisconsin Glaciation showing the location of the ice lobes. Ice extended south of the Cypress Hills to the limit of the Late Wisconsin Glaciation in Montana. Letter code in Fig. 6.2.

◌	ice covered plateaus
·•·•·•·•·	erratics limit
～	ice margin
////	coalescence area
▨	ice
xxxxxx	end moraine
000000	ice pushed gravels
➞	ice flow direction
·~·~·~·	Battle and Lyons creeks under ice cover at this time.

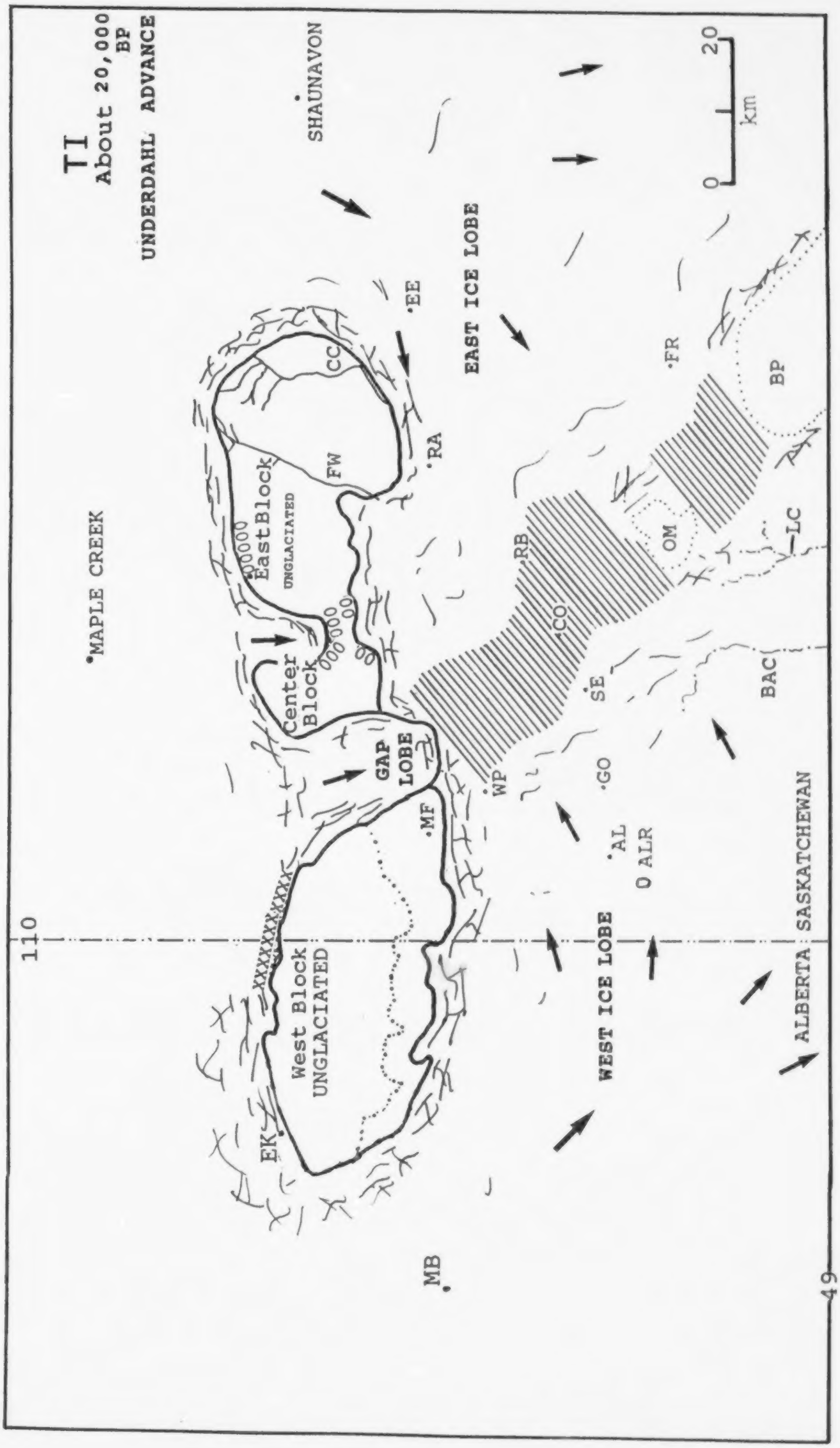

TI
About 20,000 BP
UNDERDAHL ADVANCE
SHAUNAVON
MAPLE CREEK
MB
110
49
EAST ICE LOBE
WEST ICE LOBE
East Block
UNGLACIATED
Center Block
GAP LOBE
West Block
UNGLACIATED
EK
CC
FW
EE
RA
MF
RB
CO
SE
WP
GO
AL
ALR
FR
BP
OM
LC
BAC
ALBERTA
SASKATCHEWAN
0
20
km

Figure 6.8. Time IIa, about 20,000 to 18,000 BP. The distribution of ice during the initial stages of deglaciation and the incision of the Frenchman channel. The arrows show the direction of flow along the channel. For letter code see Fig. 6.2.

·•·•·•·	erratics limit
▨	glacial lake
//////	coalescence zone
→	path of water
➞	direction of ice flow
~	ice margin
⋯	ice covered upland
— — —	uncertain margins

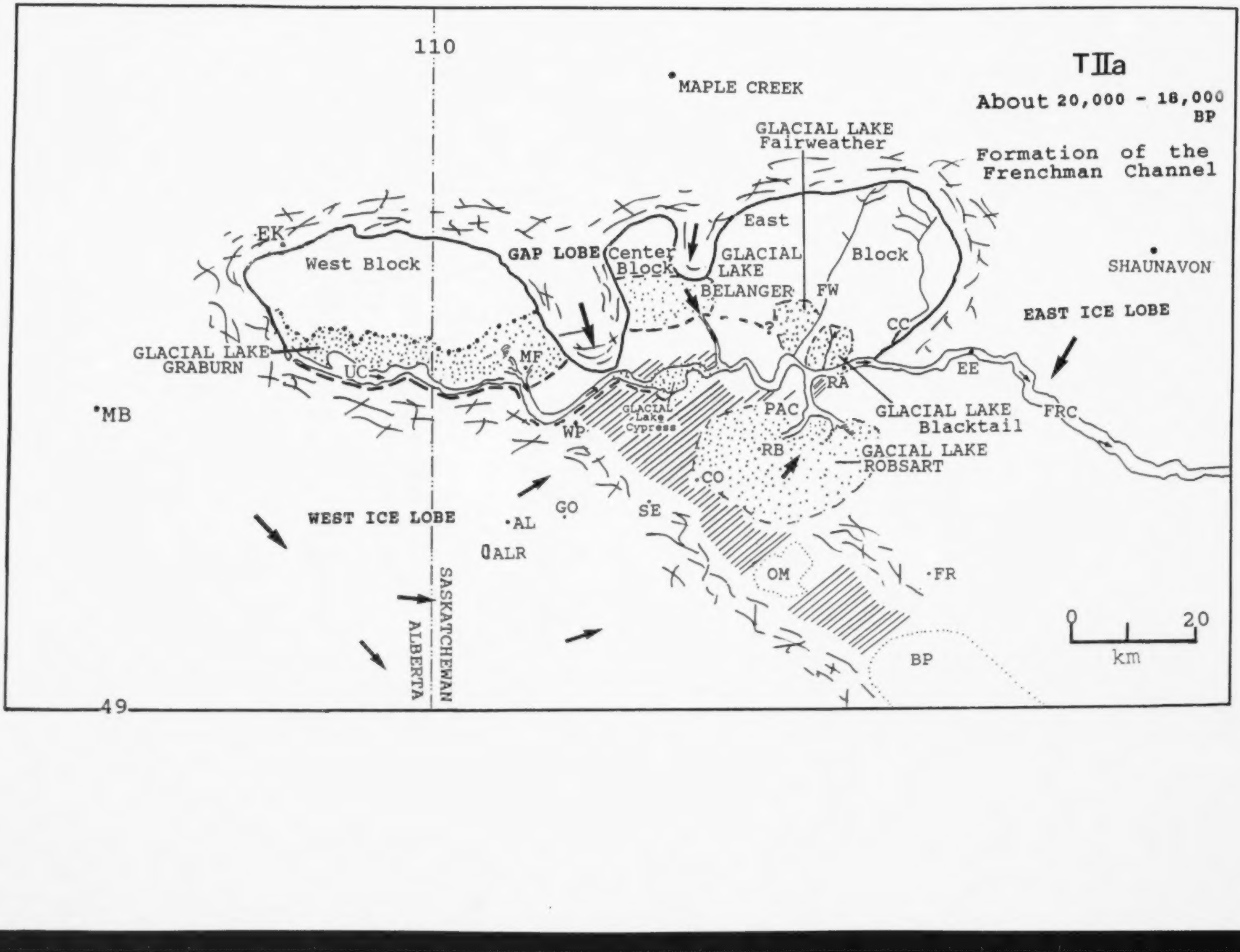
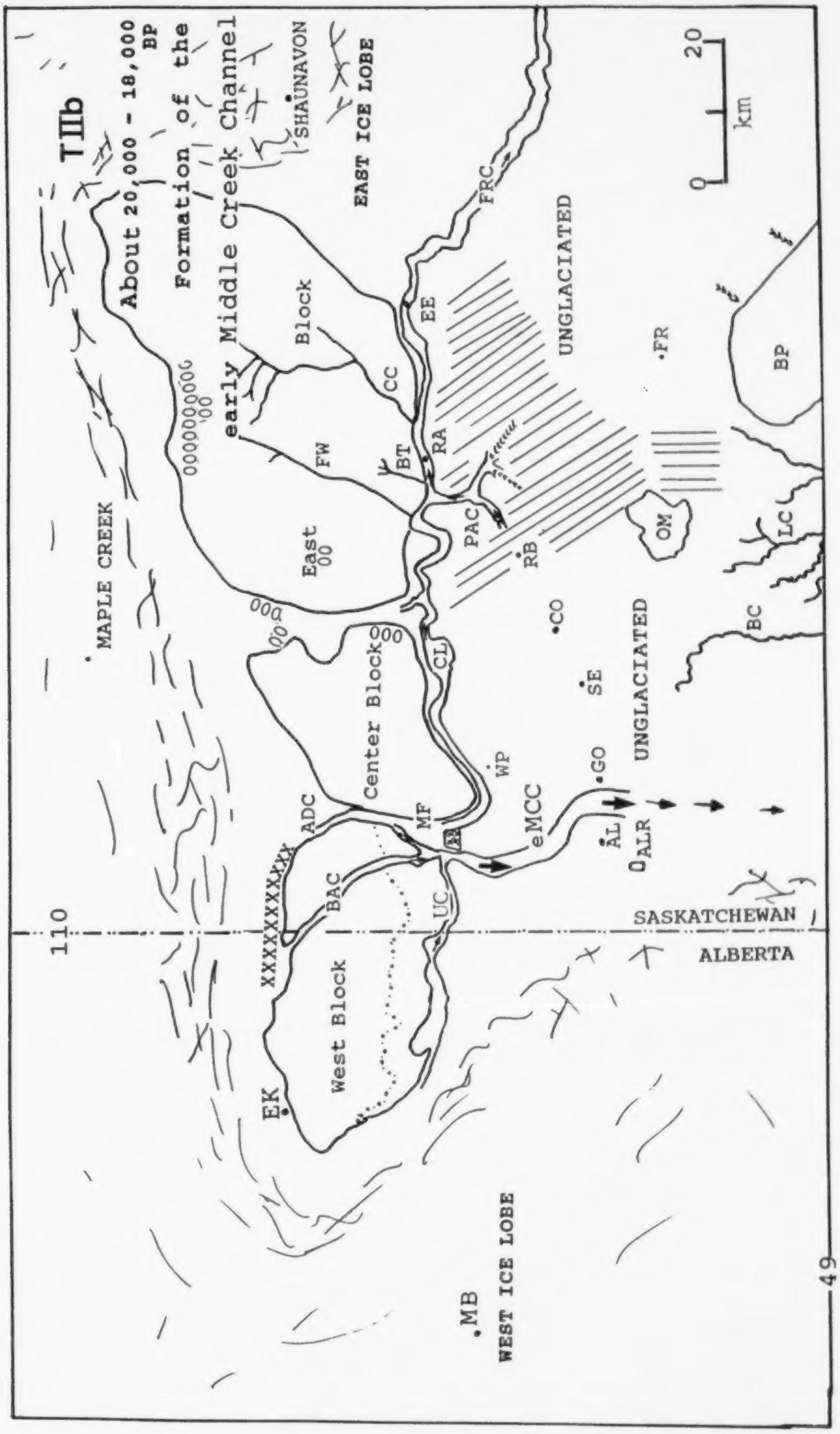

TIIa
About 20,000 - 18,000 BP
Formation of the Frenchman Channel
110
49
MAPLE CREEK
SHAUNAVON
GLACIAL LAKE Fairweather
East Block
West Block
GAP LOBE
Center Block
GLACIAL LAKE BELANGER
FW
CC
EAST ICE LOBE
EE
FRC
RA
GLACIAL LAKE Blacktail
GACIAL LAKE ROBSART
PAC
RB
CO
GLACIAL Lake Cypress
EK
GLACIAL LAKE GRABURN
UC
MF
WP
?
MB
WEST ICE LOBE
AL
GO
SE
ALR
OM
FR
BP
SASKATCHEWAN
ALBERTA
0
20
km

Figure. 6.9. Time IIb, about 20,000 to 18,000 BP. The formation of the early Middle Creek Channel (eMCC) during retreat from the Underdahl margin. The total extent of the retreat at this time is uncertain. The glacial landforms formed at the glacial maximum are also shown. Letter code given in Fig. 6.2.

∼∼∼	ice margin
>>>>>>	eskers
xxxxx	end moraine
⧹⧹⧹⧹	isolated stagnant ice
00000	ice pushed gravels
‖‖‖	ice thrust bedrock zone
➡ →	direction of water



TIIb

About 20,000 - 18,000 BP

Formation of the early Middle Creek Channel

SHAUNAVON
EAST ICE LOBE
MAPLE CREEK
East Block
Center Block
West Block
WEST ICE LOBE
UNGLACIATED
UNGLACIATED
UNGLACIATED
FRC
EE
CC
RA
BT
FW
PAC
RB
CL
CO
SE
WP
GO
eMCC
MF
ADC
BAC
UC
AL
ALR
EK
MB
FR
BP
OM
Lc
BC
110
49
SASKATCHEWAN
ALBERTA
0
20
km

Figure 6.10. TIII, The Middle Creek Advance, about 15,000 BP. The distribution of the ice during the Middle Creek Advance. The ice reached the west arm of Cypress Lake and the Middle Creek channel and partially infilled the early Middle Creek channel. The East Lobe probably reached the north bank of the Frenchman Channel. Letter code given in Fig. 6.2.

OW	outwash
⊏⊐⊏⊐⊏⊐	buried channel
•·•·•·•·•	erratics limit
⁓	ice margin
⁓ – ⁓	uncertain ice margin
⟶	ice flow direction
≫≫≫	eskers
1	hummocky ridge in the Frenchman channel

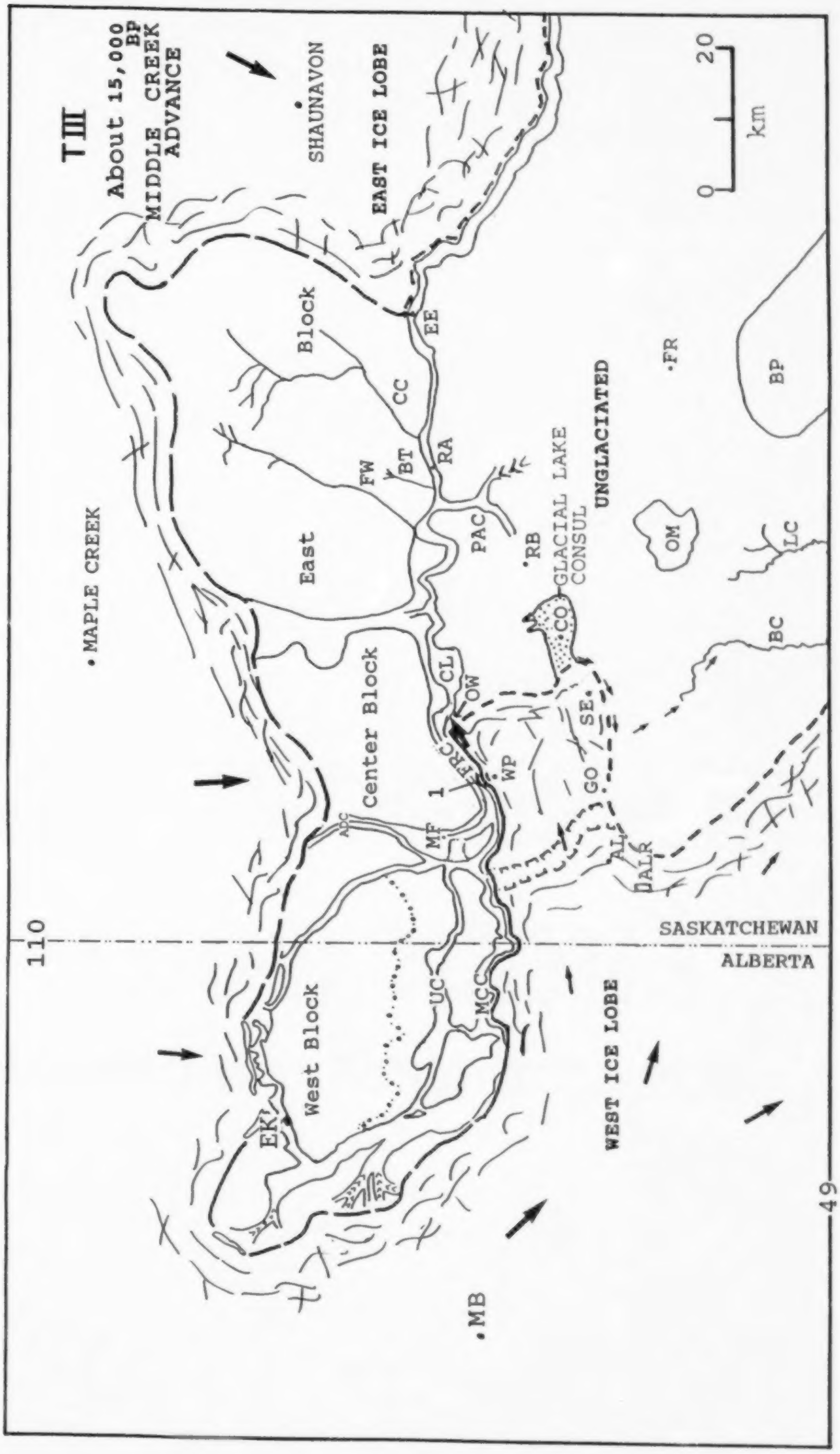

TIII
About 15,000 BP
MIDDLE CREEK ADVANCE
SHAUNAVON
EAST ICE LOBE
0
20
km
MAPLE CREEK
East
Block
FW
BT
CC
RA
EE
PAC
RB
GLACIAL LAKE CONSUL
UNGLACIATED
FR
BP
OM
Lc
CO
BC
Center Block
CL
OW
FRC
1
WP
SE
GO
ADC
MF
AL
LR
D
110
SASKATCHEWAN
ALBERTA
West Block
UC
MCC
EK
WEST ICE LOBE
49
MB

Figure 6.11. T IV, the Altawan Advance, about 15,000 to 14000 BP. The advance is restricted to the West Lobe the movements of the East Lobe at this time are uncertain. The advance partially infilled the Medicine-Lodge channel. Glacial Lake Wildhorse formed during wastage from the Altawan Margin. Letter code given in Fig. 6.2.

OW	outwash area
▭ ▭ ▭	buried channel
· • · • · • · •	erratics limit
——	ice margin
~ ~ —	uncertain ice margin
→	ice flow direction
≫≫≫	eskers
≈	Glacial Lake Wildhorse which formed during wastage of the Altawan Ice.
◠◠	Merryflat Moraine

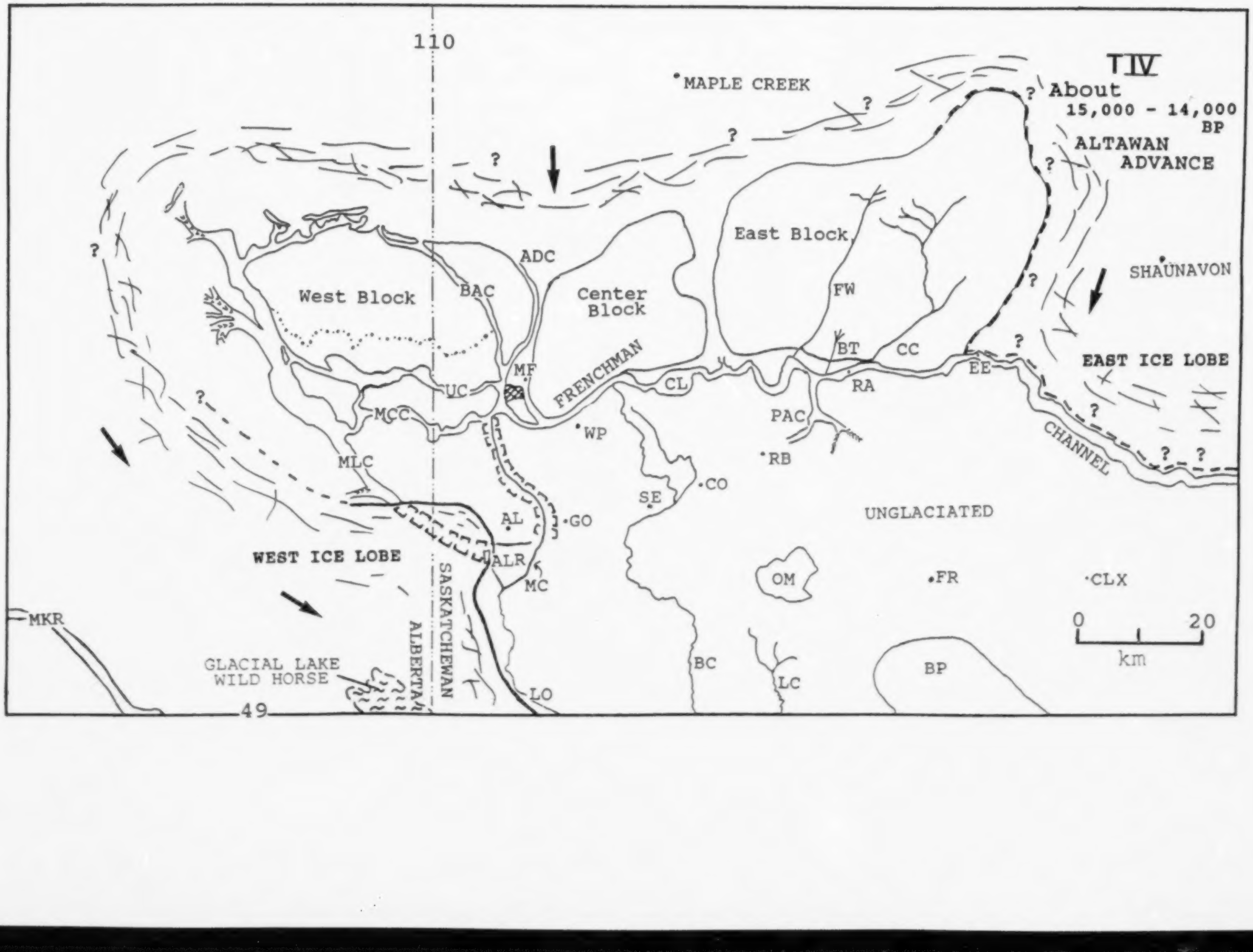

T IV
About
15,000 - 14,000
BP
ALTAWAN
ADVANCE
MAPLE CREEK
SHAUNAVON
EAST ICE LOBE
WEST ICE LOBE
110
49
West Block
Center Block
East Block
FRENCHMAN
CHANNEL
UNGLACIATED
SASKATCHEWAN
ALBERTA
GLACIAL LAKE
WILD HORSE
ADC
BAC
MF
UC
MCC
MLC
CL
FW
BT
CC
EE
RA
PAC
RB
WP
CO
SE
GO
AL
ALR
MC
LO
BC
LC
OM
FR
CLX
BP
MKR
?
0
20
km

Figure 6.12. Time V, the Pakowki Advance, about 14,000 to 13,000 BP. The Alberta portion of this map follows the margins of Westgate (1968). The ice margin north of the Cypress Hills in Saskatchewan follow a discontinuous morainal ridge. The north ice margin and the one east of the East Block is based on fieldwork and the positions in Clayton and Moran (1982). The suture zone between the West and East lobes is uncertain. The Milk River channel and Lost Creek formed at this time. Letter code in Fig. 6.2.

>>>>>>	eskers
⊏⊐⊏⊐⊏⊐	buried channel
⬭ ⬭	morainal ridge
......	erratics limit
———	ice margin
— — —	uncertain ice margin
⊥⊥⊥⊥	Westgate (1968) boundary
→	ice flow direction
▩	Merryflat Moraine

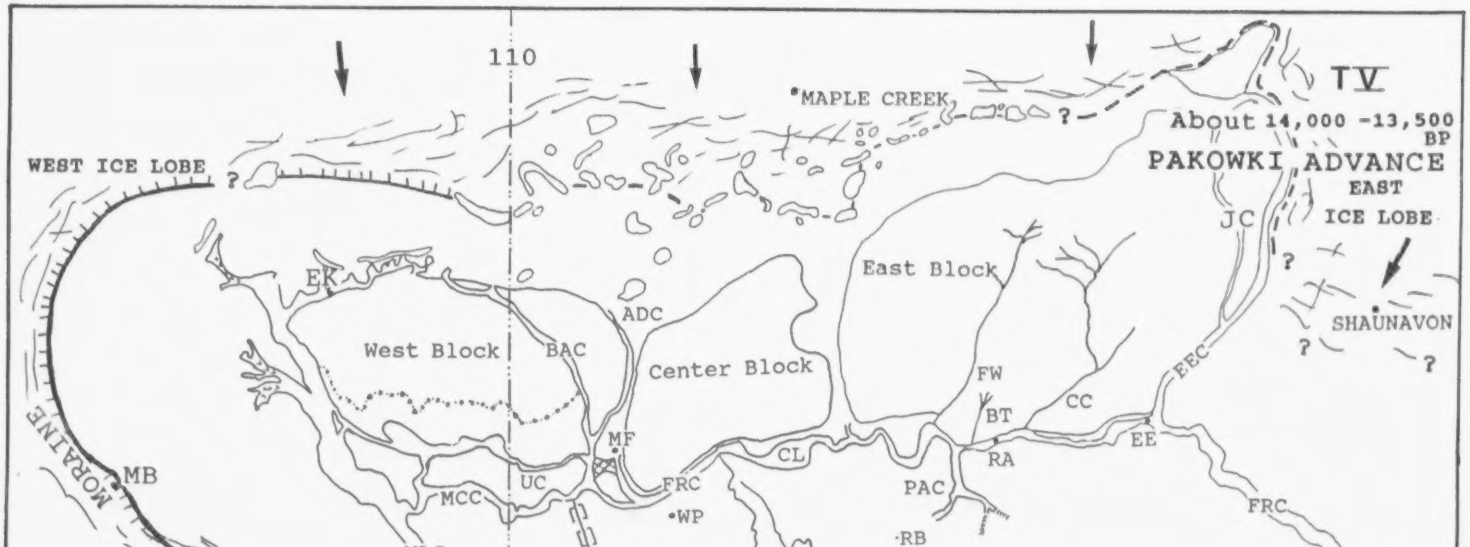
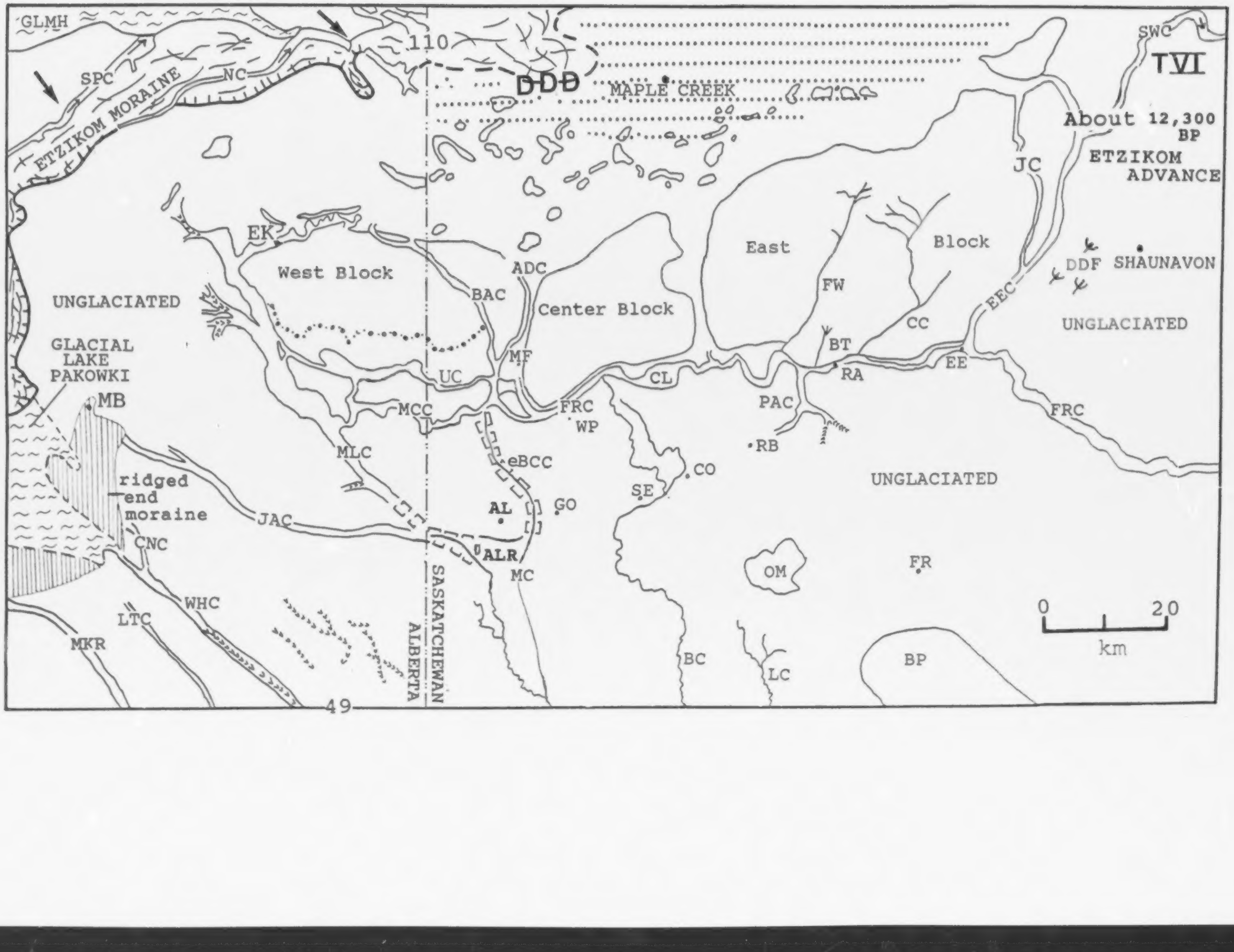

T V
About 14,000 -13,500 BP
PAKOWKI ADVANCE
EAST ICE LOBE
WEST ICE LOBE
MAPLE CREEK
SHAUNAVON
110
49
ALBERTA
SASKATCHEWAN
MORAINE
PAKOWKI
MB
EK
West Block
Center Block
East Block
UNGLACIATED
BAC
ADC
MF
UC
MCC
MLC
JAC
CNC
WHC
LTC
MKR
FRC
WP
eBCC
AL
GO
ALR
MC
LO
CL
SE
CO
BC
LC
PAC
RB
RA
BT
FW
CC
EE
EEC
JC
OM
FR
BP
0
20
km

Figure 6.13. Time VI, the Etzikom Advance, about 12,300 BP. at Time VI. The ice reached only the northwest part of the study area. The ridged end moraine marks the location of the Pakowki Advance (Westgate 1968). Glacial Lake Pakowki formed during retreat from the Pakowki Margin and drained through the Canal Coulee then the Wildhorse Channel and finally along Pakowki Channel (northeast of the study area) into the Milk River channel. Glacial Lake Medicine Hat formed during retreat from the Etzikom margin. Seven Persons Coulee and Norton Channel carried drainage to the northeast into glacial Lake Medicine Hat (Westgate 1968). The large D's represent a the approximate location of a delta which prograded into the lake impounded between the Cypress Hills and the ice to the north. Letter code in Fig 6.2.

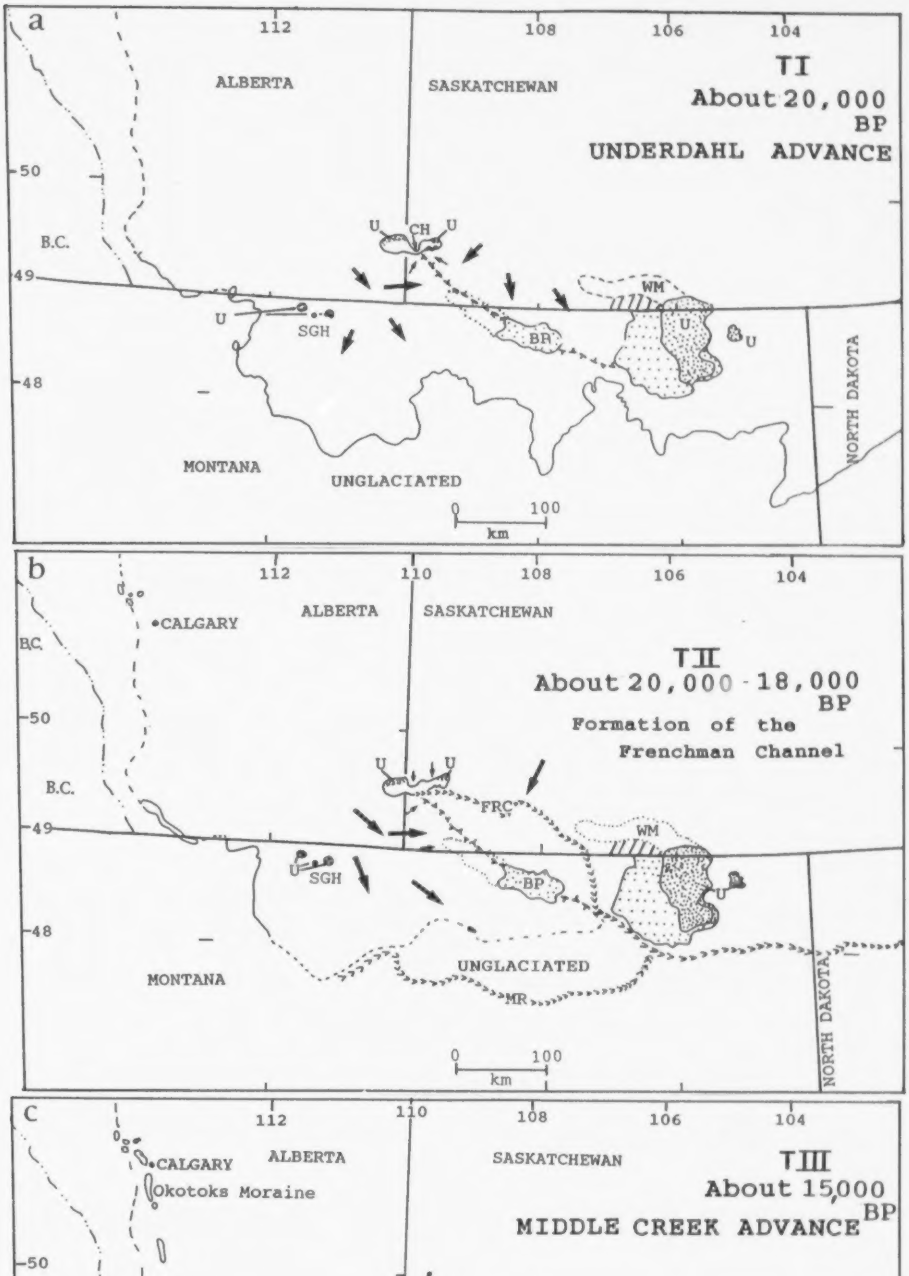
>>>>>>	eskers
⊏⊐ ⊏⊐ ⊏⊐	buried channel
->-	drumlins
⊂⊃ ⊂⊃	discontinuous moraine
⁝⁝⁝⁝	area covered by lake
D D D	inflow area
→	water flow direction
·········	erratics limit
⊥⊥⊥⊥⊥	Westgate (1968) boundary
— ~ — —	Uncertain boundary this paper

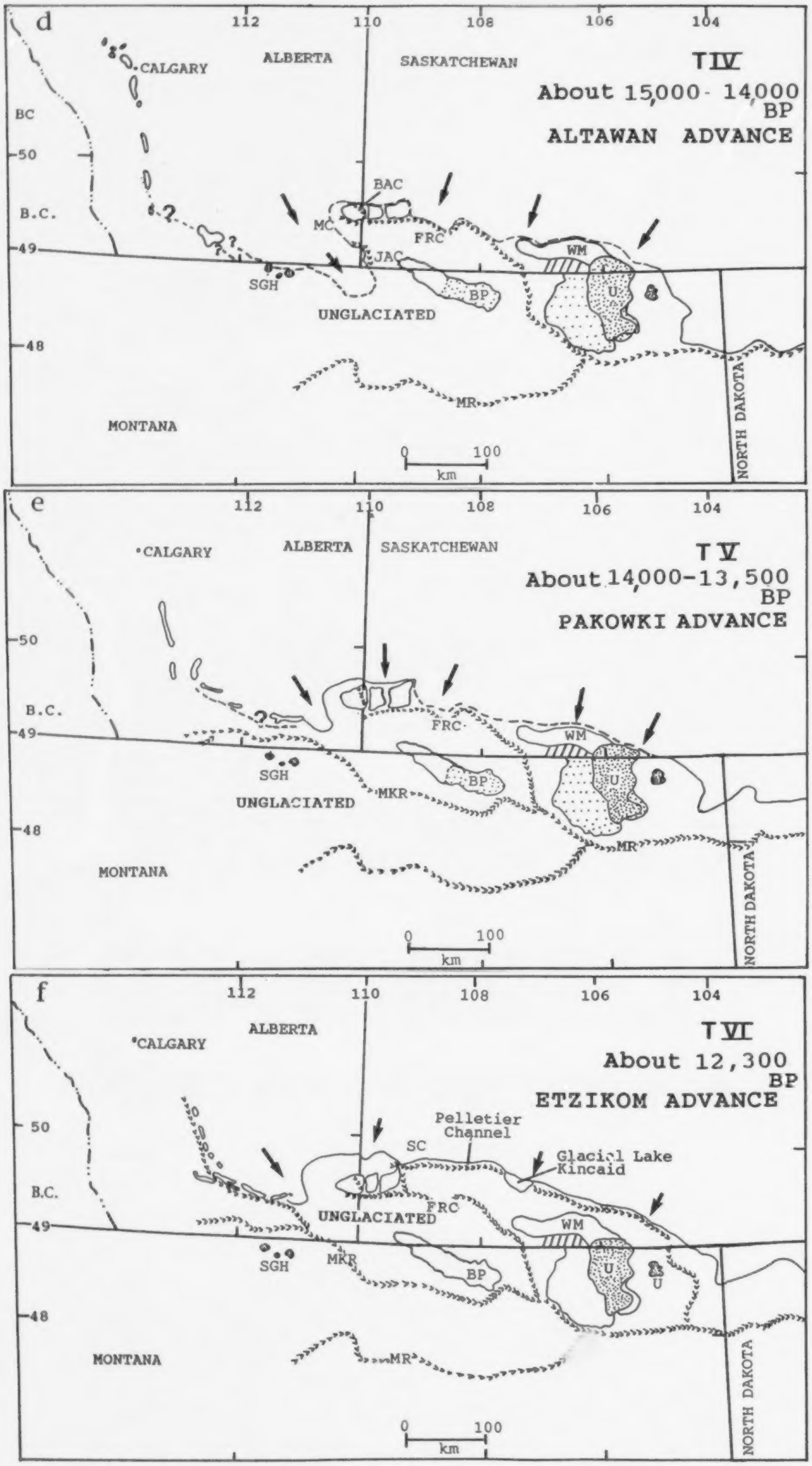


TVI
About 12,300 BP
ETZIKOM ADVANCE
GLMH
SPC
ETZIKOM MORAINE
NC
110
DDD
MAPLE CREEK
SWC
JC
EEC
DDF
SHAUNAVON
UNGLACIATED
East
Block
FW
CC
BT
RA
EE
FRC
West Block
EK
ADC
BAC
Center Block
MF
UC
CL
PAC
MCC
FRC
WP
RB
MLC
eBCC
CO
SE
GO
AL
ALR
MC
GLACIAL LAKE PAKOWKI
MB
ridged end moraine
JAC
CNC
WHC
LTC
MKR
49
SASKATCHEWAN
ALBERTA
OM
FR
BP
BC
LC
UNGLACIATED
UNGLACIATED
0
20
km

Figure 6.14. Regional distribution of the ice front at each of the six time slices described in the text. 6.14a Underdahl Advance. 6.14b Ice margin during incision of the Frenchman Channel. Ice margins out of the main study area are uncertain. 6.14c Middle Creek Advance. 6.14d Altawan Advance. 6.14e Robinson Advance. 6.14f Etzikom Advance showing Pelletier Channel and glacial Lake Kincaid. Letter Code given in Figure 6.1, 6.2. The diagrams are based on work published by Horberg (1952), Westgate (1968), Clayton and Moran (1982), Fullerton and Colton (1986), Klassen (1991, 1992) and the author's fieldwork (1988 - 1991).

SGH		Sweet Grass Hills
U	▒	never glaciated
	∼	ice boundary
	-·-·-	uncertain ice boundary
	⊥⊤⊥⊤	coalescence zone
	→	ice flow direction
	◌	ice-covered upland
	⬭	morainal ridge
	>>>>>>>	river and meltwater channels
	▨	area in southern Saskatchewan possibly covered by pre-Late Wisconsinan ice.
	⬚	pre-Late Wisconsinan affected area in Montana


a
112
108
106
104
ALBERTA
SASKATCHEWAN
TI
About 20,000 BP
UNDERDAHL ADVANCE
50
B.C.
49
48
U
CH
U
WM
U
SGH
BP
U
U
NORTH DAKOTA
MONTANA
UNGLACIATED
0
100
km
b
112
110
108
106
104
CALGARY
ALBERTA
SASKATCHEWAN
B.C.
TII
About 20,000 - 18,000 BP
Formation of the Frenchman Channel
50
B.C.
49
48
U
U
FRC
WM
U
SGH
BP
U
UNGLACIATED
MONTANA
MR
NORTH DAKOTA
0
100
km
c
112
110
108
106
104
CALGARY
ALBERTA
SASKATCHEWAN
Okotoks Moraine
TIII
About 15,000 BP
MIDDLE CREEK ADVANCE
50
B.C.
49
48
U
U
FRC
WM
U
SGH
BP
U
MONTANA
UNGLACIATED
MR
NORTH DAKOTA
0
100
km


d
112
110
108
106
104
CALGARY
ALBERTA
SASKATCHEWAN
T IV
About 15,000 - 14,000 BP
ALTAWAN ADVANCE
BC
50
B.C.
49
BAC
MC
FRC
JAC
WM
SGH
BP
U
UNGLACIATED
48
MR
MONTANA
NORTH DAKOTA
0
100
km
e
112
110
108
106
104
CALGARY
ALBERTA
SASKATCHEWAN
T V
About 14,000-13,500 BP
PAKOWKI ADVANCE
50
B.C.
49
FRC
WM
SGH
BP
U
UNGLACIATED
MKR
48
MR
MONTANA
NORTH DAKOTA
0
100
km
f
112
110
108
106
104
CALGARY
ALBERTA
T VI
About 12,300 BP
ETZIKOM ADVANCE
50
Pelletier Channel
SC
Glacial Lake Kincaid
B.C.
UNGLACIATED
FRC
49
WM
SGH
MKR
BP
U
U
48
MONTANA
MR
NORTH DAKOTA
0
100
km

References cited.

Barendregt, R.W., 1977, A detailed geomorphological survey of the Pakowki-Pinhorn area of southeastern Alberta. [Ph.D. thesis], Kingston, Ontario, Canada, Queen's University, 275 p.

Bretz, J.H. 1943. Keewatin end moraines in Alberta, Canada: Bulletin of the Geological Society of America, v. 54, p. 31-52.

Burns, J.W., Young, R.R. and Arnold, L.D., 1993, Don't look fossil gift horses in the mouth, they, too, have tongues of ice. *in* Geological Association of Canada, Mineralogical Association of Canada Program with Abstracts, Joint Annual Meeting, Edmonton 1993, p. A14.

Campbell, I., 1993, Badlands Stop 8: *in* Rutter, N.W., Campbell, I, Reasoner, M., Leslie, L., and Osborn, G., eds., Quaternary geology of the Western Canadian Prairie - Rocky Mountains, Moose Jaw to Banff Field Guide, Part II. IGCP 253 "Termination of the Pleistocene, p. 66.

Christiansen, E.A., 1968, Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 5, p. 1167-1173.

Christiansen, E.A. 1979, The Wisconsinan glaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Christiansen, E.A. 1992, Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada: an update: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Christiansen, E.A., and Sauer, E.K. 1988, Age of the Frenchman Valley and associated drift south of the Cypress Hills, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 25, p. 1703-1708.

Clayton, Lee, 1967. Stagnant-glacier features of the Missouri Coteau in North Dakota. *in* Clayton, Lee, and Freers, T.F., eds., Glacial Geology of the Missouri Coteau and adjacent areas: North Dakota Geological Survey miscellaneous Series 30, p. 25-46.

Clayton, Lee and Moran, S.R., 1982, Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 25, p. 55-82.

Clayton, Lee, Teller, J.T., and Attig, J.W., 1985, Surging of the southwestern part of the Laurentide Ice Sheet: Boreas, v. 14, p. 235-241.

Driscoll, F.G., 1980, Wastage of the Klutan ice-cored moraines, Yukon Territory, Canada: Quaternary Research, v. 14, p. 31-50.

Dyke, A.S., Dredge, L.A., and Vincent , J.-S., 1982, Configuration and dynamics of the Laurentide Ice Sheet during the late Wisconsin maximum: Geographie Physique et Quaternaire, v. 36, p. 5-14.

Dyke, A.S., and Prest, V. K., 1987a, Late Wisconsinan and Holocene history of the Laurentide ice sheet: Geographie Physique et Quaternaire, v. 41, p. 237-263.

Dyke, A.S., and Prest, V. K., 1987b, The Late Wisconsinan Maximum: Geological Survey of Canada map 1702a, scale 1:5,000,000.

Evans, D.J.A., and Campbell, I.A., 1992, Glacial and postglacial stratigraphy of Dinosaur Provincial Park and surrounding plains, southern Alberta, Canada: Quaternary Science Reviews, v. 11, p. 535-555.

Fenton, M.M., 1984, Quaternary stratigraphy of the Canadian Prairies, *in* Fulton, R.J. ed., Quaternary Stratigraphy of Canada - a Canadian Contribution to IGCP Project 24. Geological Survey of Canada, Paper 84-10. p. 57-68.

Fullerton, D.S., and Colton, R.B. 1986, Stratigraphy and correlation of the glacial deposits on the Montana Plains, *in* Richmond G.M., and Fullerton, D.S., eds., Quaternary Glaciations of the United States: Quaternary Science Reviews, v. 5, p. 69-82.

Fulton, R.J., Fenton, M.M., and Rutter, N.W., 1984, Summary of Quaternary Stratigraphy and History, western Canada, *in* Fulton, R.J. ed., Quaternary Stratigraphy of Canada - a Canadian Contribution to IGCP Project 24. Geological Survey of Canada, Paper 84-10. p. 70-83.

Hallberg, G.R., 1986, Pre-Wisconin glacial stratigraphy of the central Plains region in Iowa, Nebraska, Kansas, and Missouri: *in* Richmond G.M., and Fullerton, D.S., eds., Quaternary Glaciations of the United States: Quaternary Science Reviews, v. 5, p. 11-15.

Horberg, L., 1952, Pleistocene drift sheets in the Lethbridge region, Alberta, Canada: Journal of Geology, v. 60, p. 303-330.

Jennings, D.B., 1984, The Late Quaternary Geology of Elk Island National Park, Central Alberta, [M.S. thesis], Edmonton, Alberta Canada, University of Alberta, 197pp.

Johnston, W.A. and Wickenden, R.T. 1931, Moraines and glacial lakes in southern Saskatchewan and southern Alberta, Canada: Transactions of the Royal Society of Canada, Section 4: Geological Sciences including Mineralogy. v. 5, p. 29-44.

Klassen, R.W., 1989, Quaternary geology of the southern Canadian Interior Plains, p. 97-174. *in* Fulton, R.J., ed., Chapter 2 of the Quaternary Geology of Canada and Greenland, Geological Survey of Canada, Geology of Canada, v. 1. 839p.

Klassen, R.W., 1991, Surficial geology and drift thickness, Cypress Lake (72F) Saskatchewan: Geological Survey of Canada, Map 1766A (1:250,000).

Klassen, R.W., 1992, Nature, origin, and age relationships of landscape complexes in southwestern Saskatchewan: Geographie Physique et Quaternaire, v. 46, p. 361-368.

Klassen, R.W. and Vreeken, W.J. 1985, Quaternary geology of southwestern Saskatchewan: *in* Current Research, Part A, Geological Survey of Canada, Paper 85-1A p. 187-192.

Klassen, R.W. and Vreeken, W.J. 1987, The nature and chronological implications of surface tills and post-till sediments in the Cypress Lake area, Saskatchewan: *in*

Current Research, Part A, Geological Survey of Canada, Paper 87-A, p. 111-125.

Lemke, R.W., Laird, W.M., Tipton, M.J., Lindvall, R.M., 1965, Quaternary geology of the northern Great Plains: *in* Wright, H.E. Jr. and Frey, D.G., eds., The Quaternary geology of the United States, Princeton University Press, Princeton, New Jersey, p. 15-27.

Liverman, David, Catto, N.R., Rutter, N.W., 1989, Laurentide glaciation in west central Alberta: a single (Late Wisconsinan) event: Canadian Journal of Earth Sciences, v. 26, p. 266-274.

MacDonald, G.M., Beukens, R.P. , Kieser, W.E., and Vitt, D.H., 1987, Comparative radiocarbon dating of terrestrial plant and macrofossils and aquatic moss from the 'Ice Free Corridor' of western Canada: Geology, v. 15, p. 837-840.

McConnel, R., 1885, Report on the Cypress Hills, Wood Mountain, and adjacent country: Geological Survey of Canada, Annual Report 1-C.

Mehringer, P.J., Jr., Sheppard, J.C., and Foit, F.F., Jr., 1984, The age of Glacier Peak tephra in west-central Montana: Quaternary Research, v. 21, p. 36-41.

Moran, S.R., and Clayton, Lee, 1983, Chronology of Late Wisconsinan glaciation in middle North America -Reply: Quaternary Science Reviews, v. 1, p. 15-22.

North American Stratigraphic Code, 1983, North American Commission on Stratigraphic Nomenclature, American Association of Petroleum Geologists, v. 67, p. 841-875.

Ostrem, G., 1959, Ice melting under a thin layer of moraine and the existence of ice cores in moraine ridges: Geografiska Annaler, v. 41, p. 228-230.

Proudfoot, D.N., 1985, A lithostratigraphic and genetic study of Quaternary sediments in the vicinity of Medicine Hat, Alberta. [Ph.D. thesis], Edmonton, Alberta, University of Alberta, 345 p.

Rains, B., Shaw, J, Skoye, R., Sjogren, D., and Kvill, D., 1993, Late Wisconsinan mega-flood paths in Alberta: Geology, v. 23, p. 323-326.

Sharp, R.P. , 1949, Studies of supraglacial debris on valley glaciers: American Journal of Science, v. 247, p. 289-315.

Shetsen, I. 1984, Application of till pebble lithology to the differentiation of glacial lobes in southern Alberta: Canadian Journal of Earth Sciences, v. 21, p. 920-933.

Shetsen, I., 1987, Quaternary geology, southern Alberta: Alberta Research Council map, scale 1:500,000.

Stalker, A.MacS., 1962, Surficial geology Lethbridge (east half), Alberta: Geological Survey of Canada, Map 41-1962,

Stalker, A.MacS. 1976, Quaternary stratigraphy of the southwestern Prairies, in Quaternary stratigraphy of southern Canadian Prairies, ed. W.C. Mahaney; Dowden, Hutchinson and Ross Inc., Stroudsburg, Pennsylvania, p. 381-407.

Stalker, A.MacS. 1977, The probable extent of classical Wisconsinan ice in southern and central Alberta: Canadian Journal of Earth Sciences, v. 14, p. 2614-2619.

Stalker, A.MacS. 1980, The geology of the ice-free corridor. Canadian Journal of Anthropology, v. 1, p. 11-13.

Stalker, A.MacS., and Harrison, J.E., 1977, Quaternary glaciation of the Waterton-Castle River region of Alberta: Canadian Petroleum Geology Bulletin, v. 25, p. 882-906.

Teller, J.T., 1989, Importance of the Rossendale site in establishing a deglacial chronology along the southwestern margin of the Laurentide Ice Sheet: Quaternary Research, v. 32, p. 12-23.

Vreeken, W.J. 1986, Quaternary events in the Elkwater area of southeastern Alberta: Canadian Journal of Earth Sciences, v. 23, p. 2024-2038.

Vreeken, W.J., 1989, Late Quaternay events in the Lethbrideg area of southeastern Alberta: Canadian Journal of Earth Sciences, v. 26, p. 551-560.

Vreeken, W.J. 1991, The southern Swift Current Plateau (Saskatchewan): A sub-glacial meltwater erosion surface: Geological Association of Canada Annual Meeting, Toronto, Program with Abstracts. v. 16, p. A129.

Vreeken, W.J., and Westgate, J.A., 1992, Miocene tephra beds in the Cypress Hills of Saskatchewan, Canada: Canadian Journal of Earth Sciences, v.29, p. 48-51.

Watson, R.A., 1980, Landform development on moraines of the Klutan Glacier, Yukon Territory, Canada: Quaternary Research, v. 14, p. 50-60.

Westgate, J.A., 1965a, The surficial geology of the Cypress Hills area, Alberta: Alberta Research Council Preliminary Report 65-2, 6p.

Westgate, J.A., 1965b, The Pleistocene stratigraphy of the Foremost-Cypress Hills area Alberta, *in* Alberta Society of Petroleum Geologists, 15th Annual Field Conference, Guidebook, Part 1, Cypress Hills Plateau, p 85-111.

Westgate, J.A., 1968, Surficial geology of the Foremost - Cypress Hills area, Alberta: Research Council of Alberta Bulletin, 22. 122 pp.

Westgate, J.A., 1972, The Cypress Hills. In Quaternary geology and Geomorphology between Winnipeg and the Rocky Mountains. International Geological Congress, 24th Session Montreal Quebec, Guidebook for Excursion C-22 p. 50-62.

Westgate, J.A. and Evans, M.E., 1978, Compositional variability of Glacier Peak tephra and its stratigraphic significance: Canadian Journal of Earth Sciences, v. 15, p. 1554-1567.

Whitaker, S.H. 1965, Geology of the Wood Mountain Area (72G) Saskatchewan. [Ph.D. thesis], Urbana, Il., University of Illinois.

Young, R. R., Burns, J.A., Rains, R.B., 1989, A re-evaluation of the age of the Empress Formation gravels and Sands in the Beverly Valley, central Alberta. *in*

CANQUA 1989, Late glacial and post-glacial processes and environments in montane and adjacent areas: Programs and Abstracts. p. 50.

Young, R. R., Burns, J.A., Rains, R.B., Smith, D.G., 1993, The Late Wisconsinan Laurentide Ice sheet and its relationship to a theoretical "ice-free" corridor: *in* Program with abstracts, Geological Association of Canada, Joint Annual Meeting May 17-19, 1993, Edmonton, Alberta, p. A-113.

Young, R. R., Burns, J.A., Smith, D.G., Arnold, L.D., and Rains, R.B., 1994, A single, late Wisconsin, Laurentide glaciation, Edmonton area and southwestern Alberta: Geology, v. 22, p. 683-686.

Chapter 7 **Summary and conclusions**

This investigation into the age and genesis of glacial and glacigenic deposits on and south of the Cypress Hills of Saskatchewan has identified a widespread, thick (over 10 m) meltout till, covered by a glacigenic subaerial debris-flow assemblage along the border between Saskatchewan and Montana. The widespread distribution of the meltout till differs from the restricted meltout till distribution hypothesized by Paul and Eyles (1990). This departure may reflect several fundamental properties of the area. (1) The ice from which the meltout till was derived must have contained abundant debris (basal and englacial) to enable small features such as sand lenses and thin bedrock slabs next to them to be preserved intact. (2) The preservation of the meltout till was ensured by its position beneath the glacigenic subaerial debris-flow assemblage of Unit 2. Unit 2 acted as an insulating blanket, slowing the rate of melting of the buried ice. This enabled the underlying meltout deposit to become consolidated as it formed, thereby reducing the total amount of deformation in the meltout till. (3) The maintenance of meltwater drainage away from the ice front prevented the formation of large ice-contact lakes where sedimentation would have been dominated by glaciolacustrine depositional processes such as those described by Evenson et al. (1977), Gibbard (1980), Dreimanis (1982), and Kulig (1985).

Pleistocene glacigenic deposits exposed in the Canal, Gilchrist, and Cypress Lake sections, on the flanks of the East and Center blocks, differ markedly from the meltout till described from the Battle and Lyons creek sections. The deposits, on or near the south flanks of the East and Center blocks, have numerous diamicton beds intercalated with sand, silt, and clay laminae and other beds of glaciolacustrine origin. The encased diamicton beds therefore were also deposited in a glaciolacustrine setting. The difference in deposit type between the two areas, reflects the impounding of ice-contact lakes by the ice lobes that surrounded the Cypress Hills at the maximum of the Late Wisconsin glaciation. The coalescence of the West, Gap, and East lobes at this time, prevented meltwater from draining away from the Cypress Hills which led to the formation of the lakes. Deposition in these lakes followed the processes described by Evenson et al. (1977), Gibbard (1980), Dreimanis (1982), and Kulig (1985). The deposition of this sediment assemblage supports a conclusion reached in Chapter 2 that the maintenance of proglacial drainage has an important affect on the glacigenic processes active along an ice front and the deposits that are emplaced there.

Coalescence of the ice lobes requires redrawing of the ice lobe distribution patterns as shown by Christiansen (1979), Clayton and Moran (1982), Dyke and Prest (1987a, b), and described by Christiansen and Sauer (1988). In the new model the East Lobe extends eastward and southward to the Center Block and central Montana, respectively. The West Lobe is restricted to the area west and south of the Old Man On His Back Plateau and extends into western Montana. The Gap Lobe is confined to the area of the Center Block north of the Frenchman channel and did not continue into Montana as shown by Dyke and Prest (1987a).

This reconfiguring of the ice-lobe positions requires a rethinking of the origin of the Frenchman channel, previously interpreted as an interlobate channel gradually incised by meltwater draining between the East and West lobes along the flanks of the Cypress Hills and Wood Mountain Upland (Christiansen 1979; Clayton and Moran 1982; Dyke and Prest 1989b; Christiansen and Sauer 1988). The lack of slip-off slopes, the paucity of sorted sediment within the channel, its abrupt initiation at Merryflat, and the trench-like cross-sectional profile of the channel, match features characteristic of spillways incised by the catastrophic release of glacial lakes (Kehew 1982; Kehew and Lord 1986, 1987). Glacial lakes Belanger and Cypress, identified from the glaciolacustrine diamicton complexes observed on and near the southern slopes of the Cypress Hills (Chapter 3, this volume), were east of the head of the channel. The outburst of another glacial lake farther west is therefore needed to incise the initial reach of the channel (from Merryflat to Cypress Lake). This was Lake Graburn. Rafting of debris on icebergs, across these lakes explains the distribution of erratics beyond the limit of the last ice. Another large glacial lake, Lake Robsart, formed in the coalescence zone between the East and West lobes as indicated by the hummocky lake sediment and diamicton (Unit Mx, Klassen 1991, 1992) in this area. Lake Robsart drained northwards along Palisades Coulee and joined the Frenchman channel 3 km west of Ravenscrag. Two additional small glacial lakes, Lake Blacktail, and Lake Fairwell were probably impounded on the south flank of the East Block above Ravenscrag.

Formation of the Frenchman channel was initiated by the release of the water impounded in Lake Graburn. This triggered the subsequent outbursts of Cypress, Belanger, Robsart, Fairwell, and Blacktail lakes. At Eastend, the water turned southwards, perhaps following the course of an earlier supraglacial meltwater channel. From there it flowed along the south flank of the Wood Mountain Upland and continued southwards until it joined the Milk River in Montana.

Genetic interpretation of glacigenic sediments in the study area followed detailed section description and sedimentological analysis. The field investigations were supplemented by thin-section analysis. The thin-sections from the Saskatchewan study area were compared to thin-sections from glacigenic diamicton units from central Alberta. The comparison revealed that the subglacial meltout till had well to strongly developed skelsepic and lattisepic plasma fabric (see Brewer 1976 and Brewer and Sleeman 1988 for definitions), with clean grain boundaries, whereas debris-flow sediment has insepic plasma fabric and obscured grain boundaries. Besides plasma fabric, the thin-sections allowed detailed examination of several varieties of contacts and intraclasts. This supplemented and confirmed interpretations made from sedimentary features examined in the exposures.

The fieldwork carried out in Saskatchewan and Alberta provided sufficient information to construct a new glaciation chronology of the Cypress Hills of Saskatchewan and Alberta. The glaciation chronology was divided into six unequal time slices that can be broadly compared to those used by Christiansen (1979), Clayton and Moran (1982) and Dyke and Prest (1987b). During Time 1, around the Late Wisconsinan maximum, about 20,000 BP, the ice had surrounded the Cypress Hills and had extended into Montana. This is the equivalent to Phase D of Clayton and Moran (1982). This stage has been named the Underdahl Advance after the Underdahl channel, near its limit.

During the initial stages of retreat, the Frenchman channel was cut by the catastrophic outburst of lakes on and near the south flanks of the Cypress Hills. For this sidehill channel to be cut along the south flanks of the Center Block, East Block, and the Wood Mountain Upland, ice must have been present to the south. This southern ice must have covered the area placed by Klassen (1992) into his "pre-last advance landscape complex". This area therefore was not ice free at this time as depicted by Klassen (1992). Such extensive ice coverage to the south agrees with the Late Wisconsinan ice distribution as interpreted by Fullerton and Colton (1986). They traced their Late Wisconsinan surface till, the Fort Assiniboine Till, northwards across the border into this area of Saskatchewan. The glacial deposits exposed in sections along the Battle and Lyons creeks should be contemporaneous to the Fort Assiniboine Till.

Wastage of the ice from its maximum position continued for an unknown time interval. The areal extent of the wastage is also unknown. During this retreat, drainage was to the south along tan early stage of the Middle Creek channel. The subsequent Middle Creek Advance partially infilled this channel and continued to the west arm of Cypress Lake as indicated by an outwash fan there. During this advance, the ice reached the southern edge of the Creek marks the terminus here), deposited a small moraine in the Frenchman channel north of West Plains, and directed the Battle Creek to its present course. North of the Cypress Hills, the advance probably reached the Wild Horse (Green Lake margin) of Westgate (1968, 1972) as indicated by ice-thrust glaciolacustrine sediments at Elkwater Lake, Alberta. The lack of over-riding of the sections along the Battle and Lyons creeks, south of the Cypress Hills, indicates that the advance did not reach the Battle Creek area. The early Battle Creek cut into the underlying glacial sediments along its path and deposited fluvial sediments over the glacial ones. Since the stratified fluvial sediments are not deformed or faulted sufficient time must have passed for the complete melting of the ice from which the subglacial meltout till of Unit 1 was deposited. The interval between the Late Wisconsinan maximum Underdahl Advance and the Middle Creek advance was therefore probably several thousand years. The Middle Creek Advance broadly corresponds to Phase E of Clayton and Moran (1982) but is areally, less extensive.

During retreat from the Middle Creek margin, the Medicine Lodge channel transported meltwater around the West Block. This channel was partially infilled by the following Altawan Advance, which extended only a few kilometers into southern Saskatchewan. At Govenlock and the Altawan reservoir, ice from the Middle Creek Advance was buried by sediments derived from the Altawan Advance. The persistence of this ice indicates that the duration of this retreat was probably shorter than that between the Underdahl and Middle Creek advances. The ice limit north of the Cypress Hills is uncertain but it probably did not reach above 1060 m. The Altawan Advance has no direct equivalent in either the Westgate (1968, 1972) or Clayton and Moran (1982) chronologies.

The Altawan Advance was followed in Alberta by the Pakowki Advance as described by Westgate (1968). North of the Cypress Hills in Saskatchewan, a discontinuous ridge of glacial sediment extends eastward for about 60 km. This ridge connects to a similar more discontinuous ridge system in Alberta that marks the limit of the Pakowki Advance (Westgate 1968) and therefore is the most likely limit of the Pakowki Advance in Saskatchewan. Near the end of this time period, the Eastend Coulee was cut along the east margin of the East Block by drainage of a large lake that covered much of the area north of the Cypress Hills (Klassen 1992).

The Etzikom Advance position corresponds to that drawn by Westgate (1968), Christiansen (1979), and Clayton and Moran (1982). This advance ended when meltwater impounded in the large lake north of the Cypress Hills drained north and eastwards along Swift Current Creek and the Pelletier channel.

The pattern of ice retreat differs markedly between the East and West lobes. The East Lobe appears to have been relatively stable during much of the time of deglaciation and appears to have withdrawn only a few tens of kilometers. The West Lobe during the same time underwent several extensive advances and retreats. The glaciation pattern of the West Lobe is more similar to the surging lobes described by Clayton et al. (1985). The different behavior of the East and West lobes may indicate that they were not coupled. This may reflect a difference in ice source for each lobe. The West Lobe likely originated from an ice divide east of Great Slave Lake in the Northwest Territories. Whereas, the East Lobe and most of the ice north of the Cypress Hills was derived from an eastern Laurentide ice center (Shetsen 1984). The formation of a western ice divide (see Dyke et al. 1982) that had not formed during earlier glacial episodes may explain why the Late Wisconsinan ice was the most extensive ice to affect the Cypress Hills. This agrees with Liverman et al. (1989) who determined that only Late Wisconsinan ice covered the Watino area of central Alberta.

References cited

Brewer, R., 1976, Fabric and mineral analysis of soils., 2nd edition. Wiley, New York, 482 pp.

Brewer, R., and Sleeman, J.R., 1988, Soil structure and fabric: CSIRO Australia, Melbourne Australia, 174p.

Christiansen, E.A., 1979, The Wisconsinan glaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Christiansen, E.A. and Sauer, E.K., 1988, Age of the Frenchman Valley and associated drift south of the Cypress Hills, Saskatchewan, Canada: Canadian Journal of Earth Sciences, v. 25, p. 1703-1708.

Clayton, Lee and Moran, S.R., 1982. Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55-82.

Clayton, Lee, Teller, J.T., and Attig, J.W., 1985, Surging of the southwestern part of the Laurentide Ice Sheet: Boreas, v. 14, p. 235-241.

Dreimanis, A., 1982. Two origins of the stratified Catfish Creek Till at Plum Point, Ontario. Boreas, v. 11, p. 173-180.

Dyke, A.S., and Prest, V. K., 1987a, Late Wisconsinan and Holocene history of the Laurentide ice sheet: Geographie Physique et Quaternaire, v. 41, p. 237-263.

Dyke, A.S., and Prest, V. K., 1987b, The Late Wisconsinan Maximum: Geological Survey of Canada map 1702a, scale 1:5,000,000.

Evenson, E.D., Dreimanis, A., Newsome, J.W., 1977, Subaquatic flow till: a new interpretation for the genesis of some laminated till deposits: Boreas. v. 6, p. 115-133.

Fullerton, D.S., and Colton, R.B. 1986, Stratigraphy and correlation of the glacial deposits on the Montana Plains. *in* Richmond, G.M., and Fullerton, D.S.., eds., Quaternary Glaciations of the United States: Quaternary Science Reviews, v. 5, p. 69-82

Gibbard, P. , 1980, The origin of the Catfish Creek Till by basal melting: Boreas, v. 9, p. 71-80.

Kehew, A.E., 1982, Catastrophic flood hypothesis for the origin of the Souris spillway, Saskatchewan and North Dakota: Geological Society of America Bulletin, v. 93, p. 1053-1058.

Kehew, A.E. and Lord, M.L., 1986, Origin and large-scale erosional features of glacial-lake spillways in the Northern Great Plains: Geological Society of America Bulletin, v. 97, p. 162-177.

Kehew, A.E., and Lord, M.L., 1987, Glacial-lake outbursts along the mid-continent margins of the Laurentide ice-sheet. *in*: Mayer, L. and Nash, D., eds., Catastrophic Flooding, Allen and Unwin, Boston, p. 95-120.

Klassen, R.W., 1991, Surficial geology and drift thickness, Cypress Lake (72F) Saskatchewan: Geological Survey of Canada, Map 1766A (1:250,000).

Klassen, R.W., 1992, Nature, origin, and age relationships of landscape complexes in southwestern Saskatchewan: Geographie Physique et Quaternaire, v. 46, p. 361-368.

Kulig, J.J., 1985, A sedimentation model for the deposition of glacigenic deposits in central Alberta. [M.S. thesis], Edmonton, Alberta, Canada, University of Alberta, 245pp.

Liverman, David, Catto, N.W., Rutter, N.W., 1989, Laurentide glaciation in west central Alberta: a single (Late Wisconsinan) event: Canadian Journal of Earth Sciences, v. 26, p. 266-274.

Paul, M.A., and Eyles, N., 1990, Constraints on the preservation of diamict facies (melt-out tills) at the margins of stagnant glaciers: Quaternary Science Reviews: v. 9, p. 51-69,

Shetsen, I. 1984, Application of till pebble lithology to the differentiation of glacial lobes in southern Alberta: Canadian Journal of Earth Sciences, v. 21, p. 920-933.

Westgate, J.A., 1968, Surficial geology of the Foremost - Cypress Hills area, Alberta: Research Council of Alberta Bulletin 22, 122 pp.

Westgate, J.A., 1972, The Cypress Hills. *in* Quaternary geology and Geomorphology between Winnipeg and the Rocky Mountains. International Geological Congress, 24th Session Montreal Quebec, Guidebook for Excursion C-22 p. 50-62.

Appendix 1: Terminology

There are many classification schemes that have been applied to glacial and glacigenic deposits. A substantial review of the genetic terminology as applied to till is given in Dreimanis (1989). Because there is such a variety of definitions in the literature this appendix will provide definitions of the main terms as used in this thesis.

Diamicton - is a non-genetic term for a none sorted or poorly sorted noncalcareous, terrigenous rock that contains a wide range of particle sizes. Using this definition the common term stratified diamicton becomes problematical. For example if a unit is made up of a number of diamicton layers interbedded with well-sorted sand or silt interlayers the well-sorted interlayers should be considered separately as they do not strictly fit the definition of being poorly sorted. It is better to describe the unit as an assemblage or complex than as a single stratified diamicton.

Till has been defined in many different ways. The greatest difficulty in assigning a definition to a term is that the glacial sedimentary environment is characterized by numerous gradual transitions from one sediment type to another. The problem will always exist as to where to place the boundaries between such sediment types. Till is defined for the purposes of this thesis as a sediment that has been deposited directly by glacier ice and has not undergone extensive subsequent disaggregation and remobilization. Using this definition restricts the use of the word to what are often considered primary tills, i.e., varieties of meltout, lodgement. Shaw (1982) produced a chart for the classification of terrestrial tills in which the position of transport, position of formation, and the process of formation are the important factors. Debris could be carried in supraglacial, englacial or basal positions. It can be released in proglacial, lateral, supraglacial, englacial, or subglacial positions. Finally it can form by lowering, flow, sublimation, meltout, or lodgement. In this thesis flow tills are not considered to be tills because the debris-flow sedimentation process removes the primary features inherited from the ice and mixes the sediment with material deposited by other processes, i.e., subaerial glacigenic debris-flows are intermixed with pond sediment, slope wash, and glaciofluvial sediment.

Meltout till "is deposited by slow release of glacial debris from ice that is not sliding or internally deforming" Dreimanis (1989, p. 45). Some of the features of the debris-rich ice are retained by the meltout till. There are several types of meltout tills have been named. The most commonly observed are supraglacial and subglacial meltout tills. Using the classification scheme of Shaw (1982) a case can be made for calling a meltout till, a basal meltout till if it is composed of material transported basally and if the position of formation is uncertain. This thesis has used the term subglacial meltout till, i.e., one deposited from basal ice at the bottom of an unmoving ice mass, to reduce confusion as to the exact origin of the deposit.

Lodgement till "is deposited by plastering of glacial debris from the sliding base of a moving glacier by pressure melting and/or other mechanical means." Dreimanis (1989, p. 43).

Another important component observed in the diamicton units of the study area are soft-sediment intraclasts of various types. Diamicton intraclasts are irregular to sub-rounded soft sediment clasts which range from coarse to large pebbles in size. They commonly have the same grain size distribution as the encasing diamicton but lack the very coarse sand to pebbles of the encasing diamicton. Usually diamicton intraclasts are only detectable in thin section where discontinuities and grain arrangements in the clasts differ from the surrounding diamicton. Pebble-core diamicton clasts have a pebble center encased in fine-grained diamicton (low sand content). Silt intraclasts are composed of irregular to rounded coarse sand to pebble-sized clasts of silt. Grading is observable in some of these intraclasts. The grading reflects from derivation of the intraclast from the break-up or erosion of previously deposited normally graded silt layers. In some places, diamicton beds composed of silt intraclasts in a silt matrix are found. These diamicton beds were derived from the remobilization of silt beds by slumpage and debris flow. Silt and clay couplet intraclasts are those composed of a silt and clay couplet. There is commonly a gradational contact between the silt and clay parts of the intraclast. These intraclasts were formed by the break-up of silt and clay couplet laminae and beds.

In the discussion of glaciolacustrine sedimentation two terms and the distinctions between them must be made clear. The terms are varve and rhythmite. A varve is a sedimentary bed or lamina deposited in a still body of water over a period of one year. A rhythmite is a couplet with two distinct components or a graded sequence of sediments that form a unit in a layer of rhythmically bedded sediments. There is no limit as to the thickness or complexity of the unit, and the unit does not possess any time connotation. Numerous studies have shown that more than one rhythmite can be deposited in a single year (see Lambert and Hsu, 1979; Smith and Ashley 1986) therefore when describing glaciolacustrine sediments and glaciolacustrine diamicton assemblages, the term rhythmite, as it is time independent, is preferable.

References Cited

Dreimanis, A.E., 1989, Tills: their genetic terminology and classification. *in* Goldthwait, R.P. and Matsch, C.L., eds., Genetic Classification of Glacigenic Deposits: A.A. Balkema, Rotterdam, p. 17-84.

Lambert, A., and Hsu, K.J., 1979, Non-annual cycles of varve-like sedimentation in Walensee, Switzerland: Sedimentology, v. 26, p. 453-461.

Shaw, J., 1982, Meltout till in the Edmonton area, Alberta, Canada: Canadian Journal of Earth Science, v. 9, p. 1548-1569.

Smith, D., and Ashley, G., M., 1985, Chapter 4, Proglacial lacustrine environment, *in* Ashley, G. M., Shaw, J., and Smith, N., D., eds., Glacial Sedimentary Environments. Special Publication 16, Society of Economic Paleontologists and Mineralogists, Tulsa, OK, p. 175-216.

Appendix 2: Rose Diagrams for Chapters 2 and 3.

This appendix contains bi-directional Rose diagrams for the fabrics measured in Chapters 2 and 3. Summary statistics are included beside the diagram. The Rose diagrams are bi-directional ones produced by the Rosy orientation analysis program of D.B. McEachran (1987 version). The Rose diagrams are important in showing which fabrics are multimodal. Multimodal fabrics give averaged trend directions which may not be accurate as this type of orientation analysis was developed for unimodal data. Good examples of trend averaging which have significant effects on the trend direction is diagram CS2-90. The averaged trend is 299^0 whereas the strongest primary trend is centered around 315^0. On a highly multimodal diagram the averaged trend is therefore not predictive. North is to the top of the page. Five degree counting intervals are used in most diagrams (eg. LC1-1, LC1-2), the rest (eg. BC1-1, BC1-2) use ten degree counting intervals.

Rose Diagrams for Chapter 2.

Battle Creek Section 1

BC1-1

Trend - 48.4

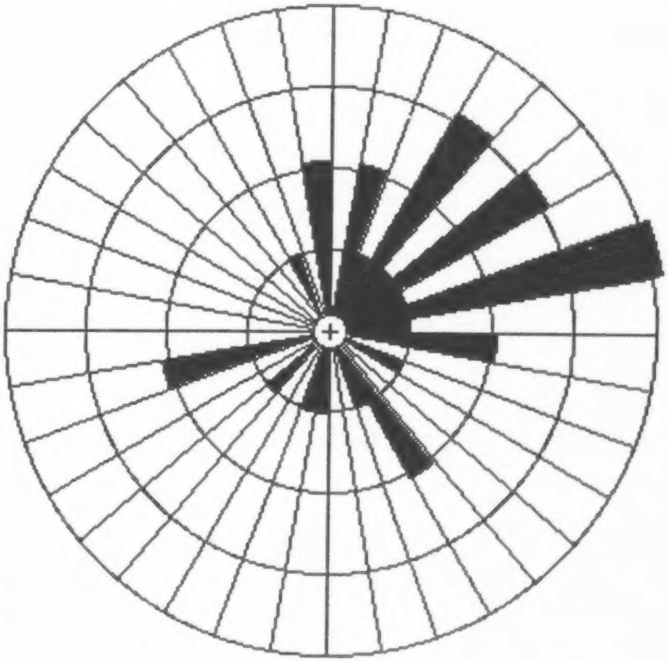
Plunge - 13.7

S_1 - .62

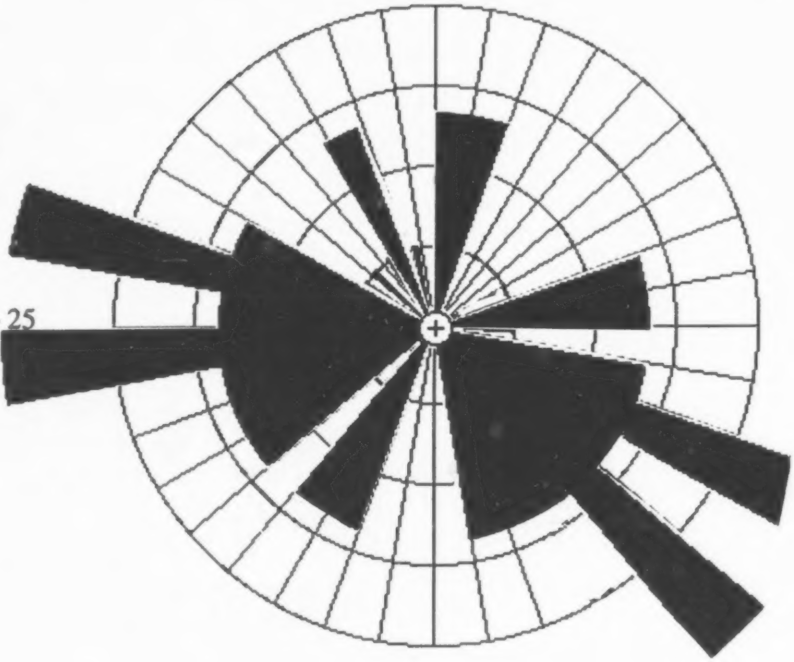
S_2 - 0.29

Clast Number - 30

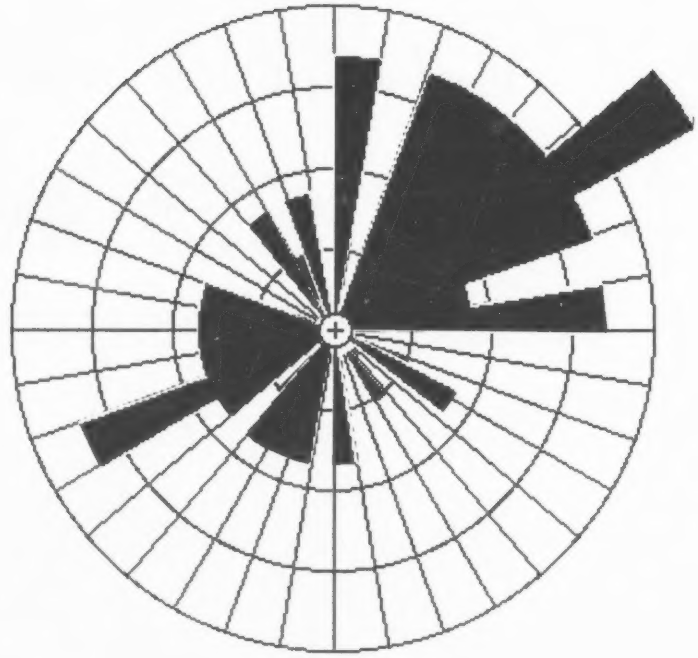
Multimodal



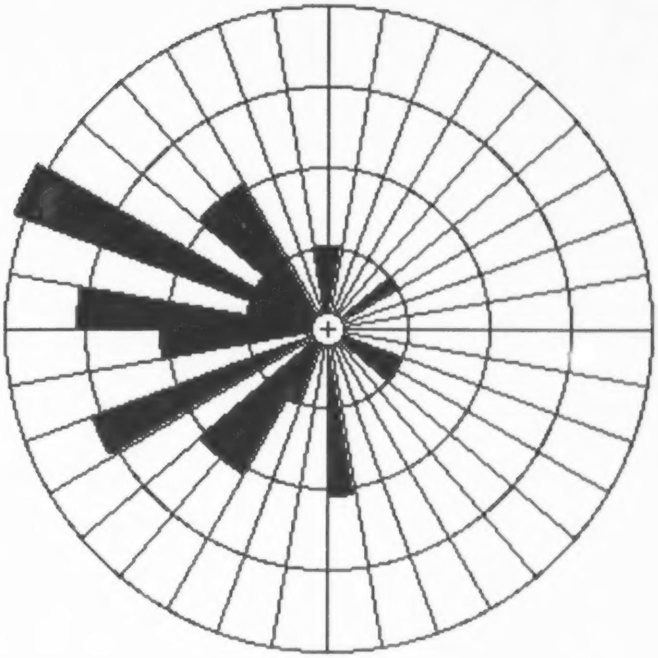
BC1-2
Trend - 282.2
Plunge - 3.5
S_1 - 0.55
S_2 - 0.28
Clast Number - 25
Multimodal



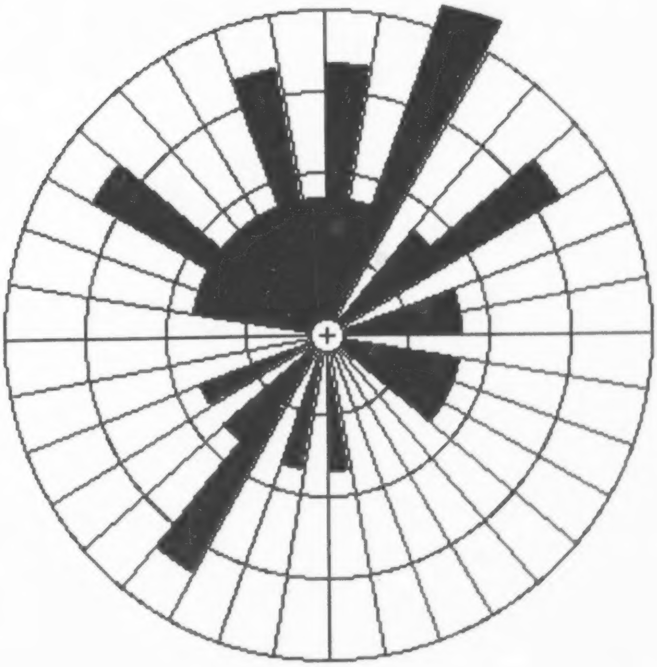
BC1-3
Trend - 48.2
Plunge - 6.8
S_1 - 0.65
S_2 - 0.24
Clast Number - 30
Multimodal

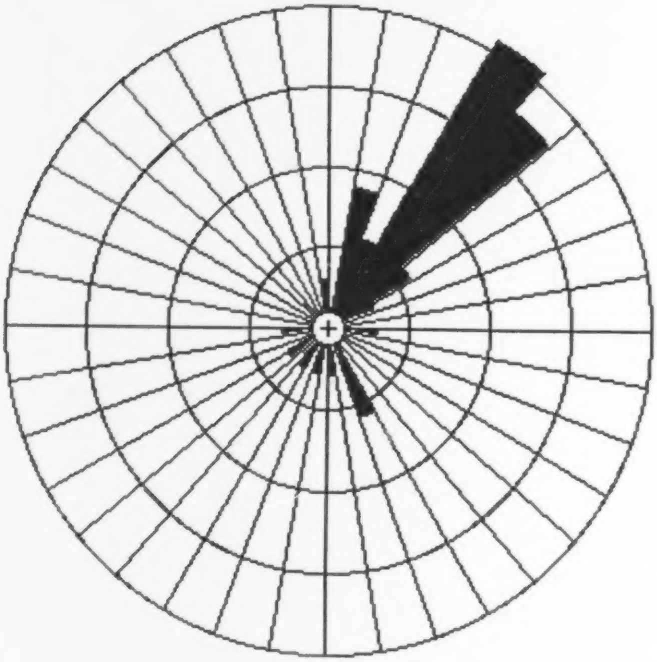


BC1-4
Trend - 277.6
Plunge - 30.6
S_1 - 0.61
S_2 - 0.30
Clast Number - 30
Multimodal

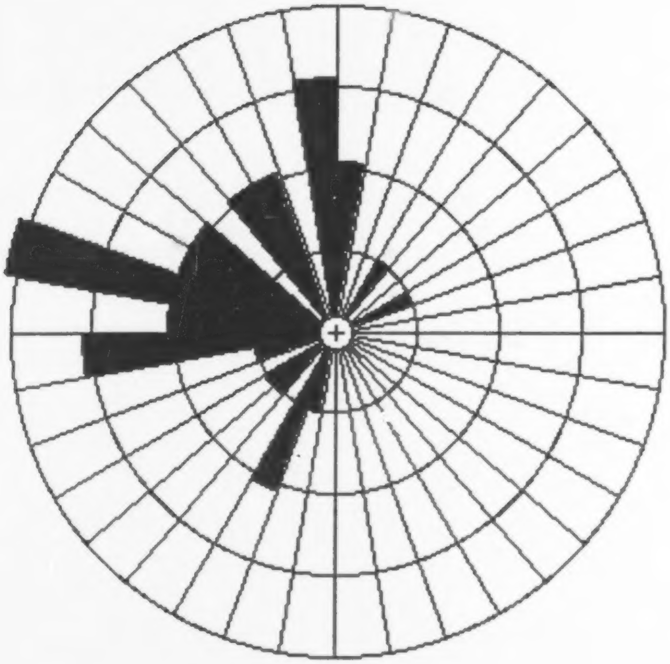


BC1-5
Trend - 5.4
Plunge - 12.8
S_1 - 0.54
S_2 - 0.40
Clast Number - 30
Multimodal



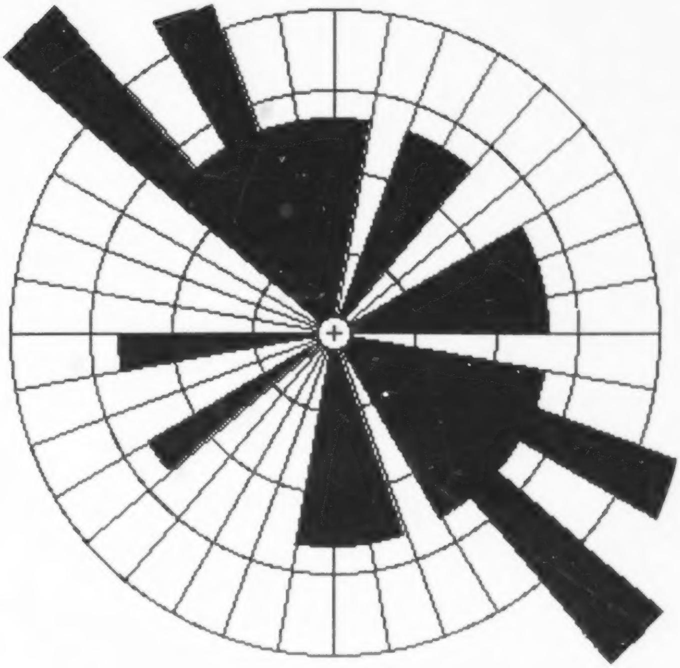


BC1-6
Trend - 305.9
Plunge - 22.3
S_1 - 0.77
S_2 - 0.15
Clast Number - 30
Unimodal

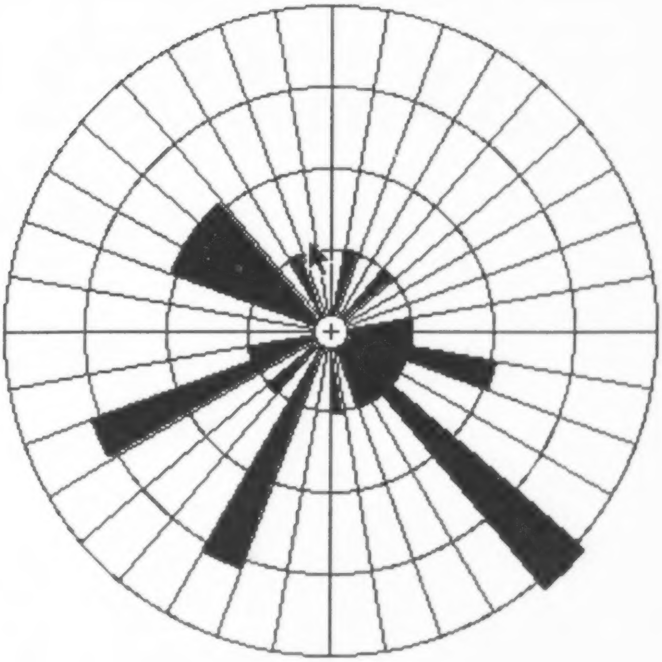


BC1-7
Trend - 288.1
Plunge - 30.0
S_1 - 0.58
S_2 - 0.34
Clast Number - 30
Multimodal

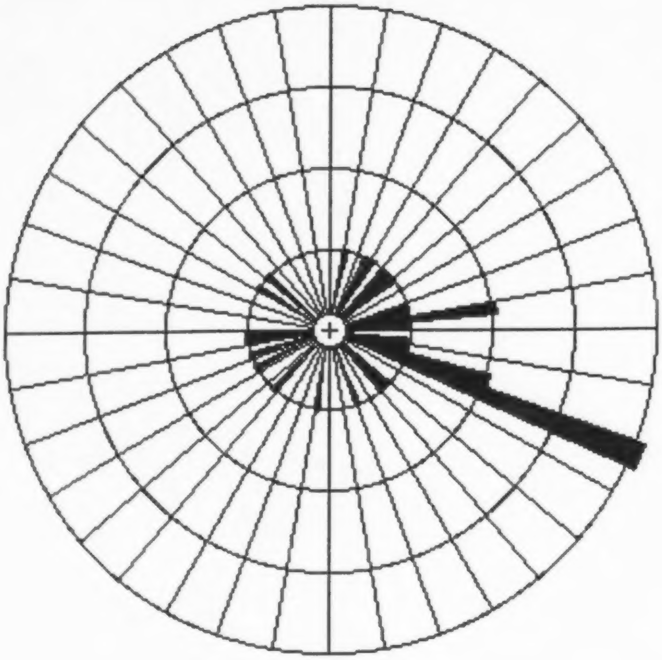
BC1-8
Unit - 2
Trend - 320.7
Plunge - 1.8
S_1 - 0.58
S_2 - 0.31
Clast Number - 25
Multimodal



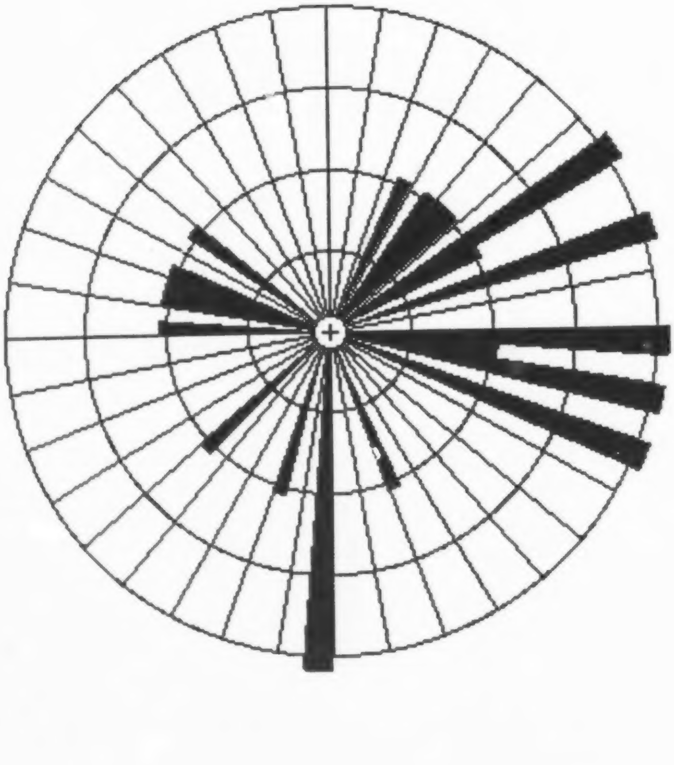
BC1-9
Unit - 2
Trend - 123.1
Plunge - 2.2
S_1 - 0.58
S_2 - 0.32
Clast Number - 30
Multimodal



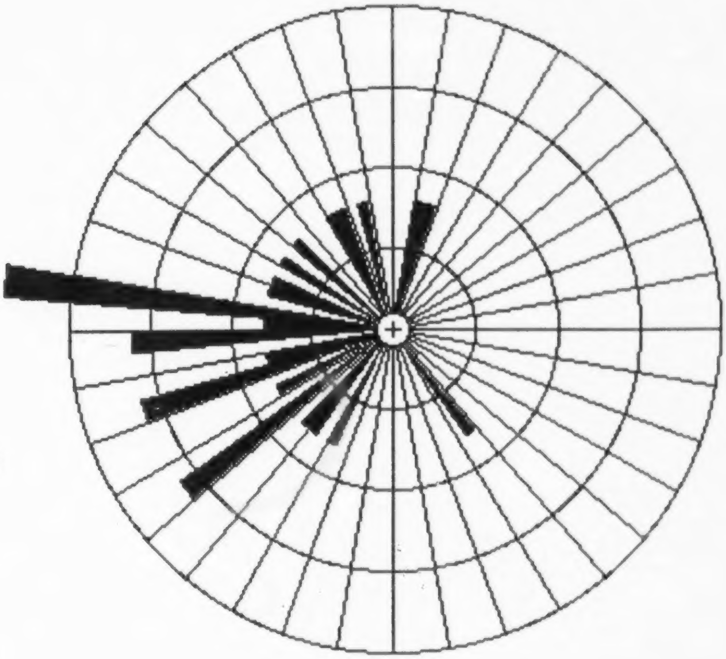
LC1-1
Unit - 1
Trend - 90.6
Plunge - 10.0
S_1 - 0.63
S_2 - 0.24
Clast Number - 30
Multimodal



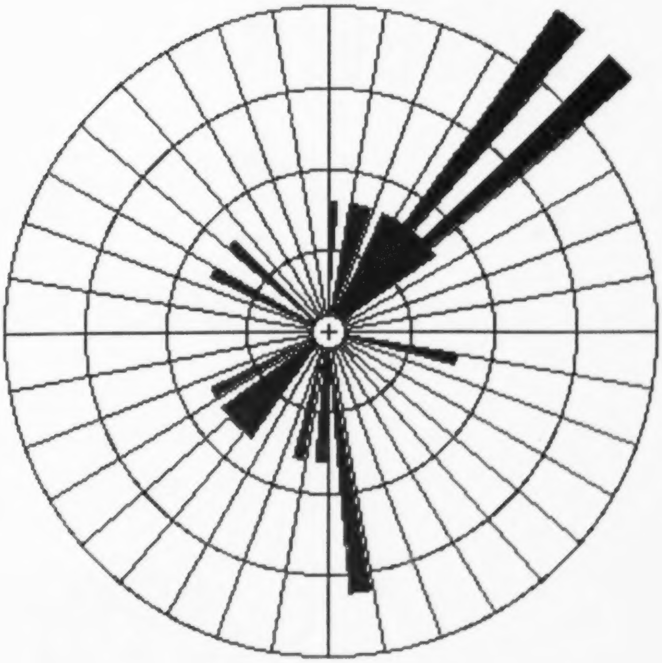
LC1-2
Unit - 1
Trend - 77.7
Plunge - 10.7
S_1 - 0.59
S_2 - 0.28
Clast Number - 26
Multimodal



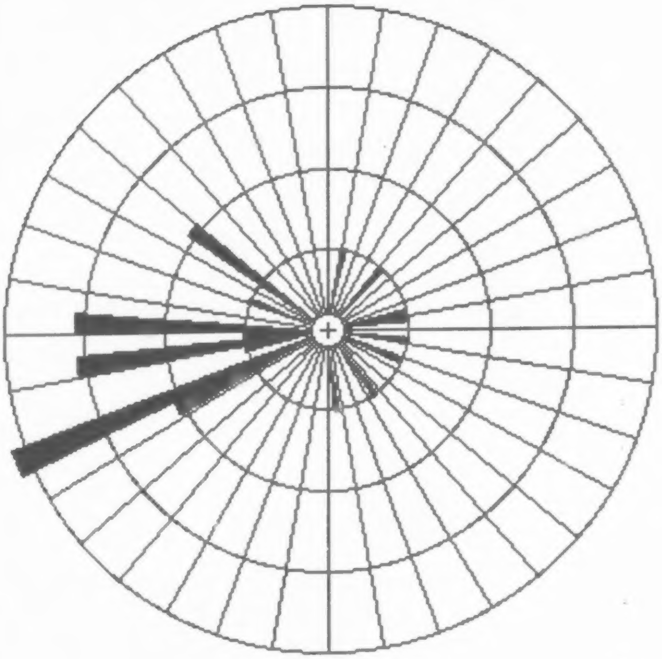
LC1-3
Unit - 1
Trend - 267.3
Plunge - 36.6
S_1 - 0.64
S_2 - 0.29
Clast Number - 25
Multimodal



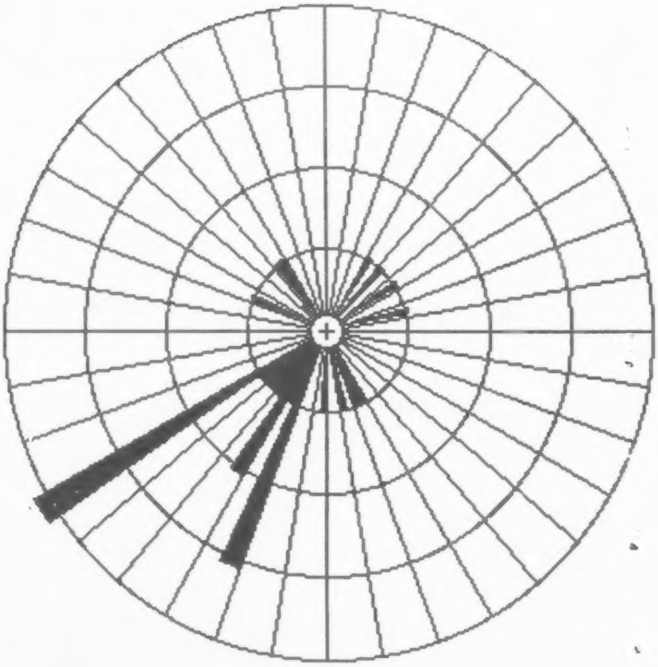
LC1-4
Unit - 1
Trend - 34.5
Plunge - 2.6
S_1 - 0.73
S_2 - 0.17
Clast Number - 25
Multimodal



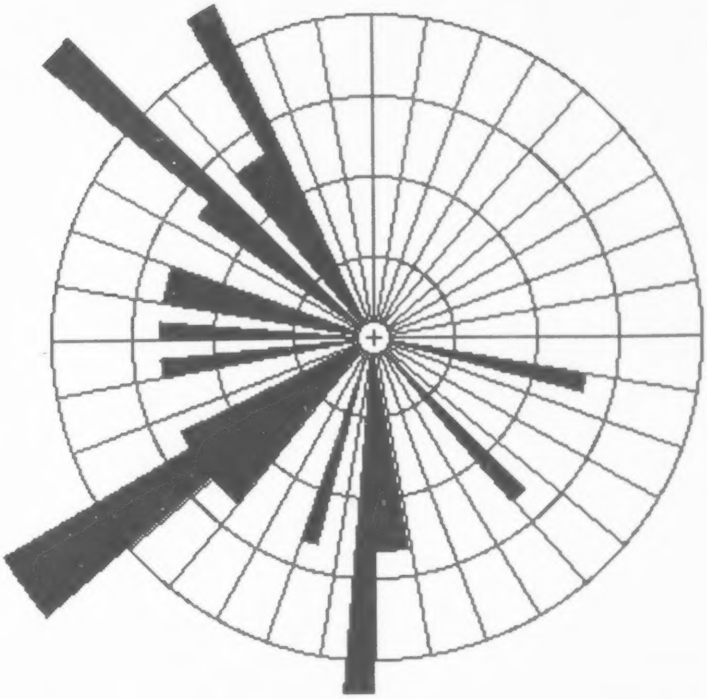
LC1-5
Unit - 1
Trend - 264.4
Plunge - 22.8
S_1 - 0.70
S_2 - 0.14
Clast Number - 25
Multimodal



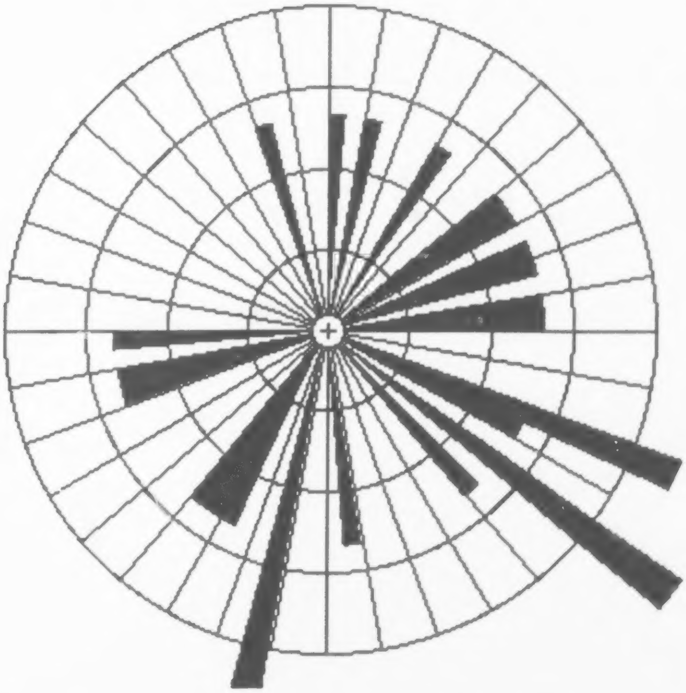
LC1-6
Unit - 1
Trend - 123.6
Plunge - 8.7
S_1 - 0.71
S_2 - 0.24
Clast Number - 25
Multimodal



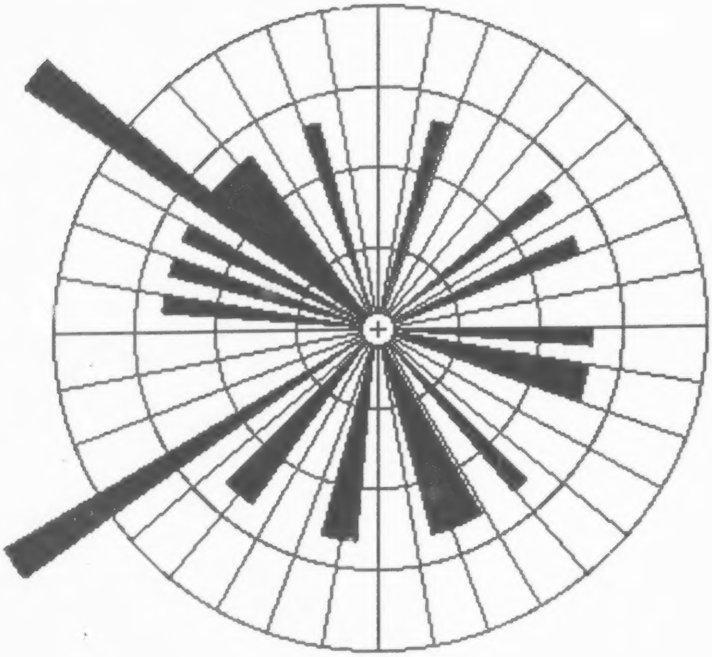
LC1-7
Unit - 1
Trend - 268.9
Plunge - 35.8
S_1 - 0.53
S_2 - 0.33
Clast Number - 25
Multimodal



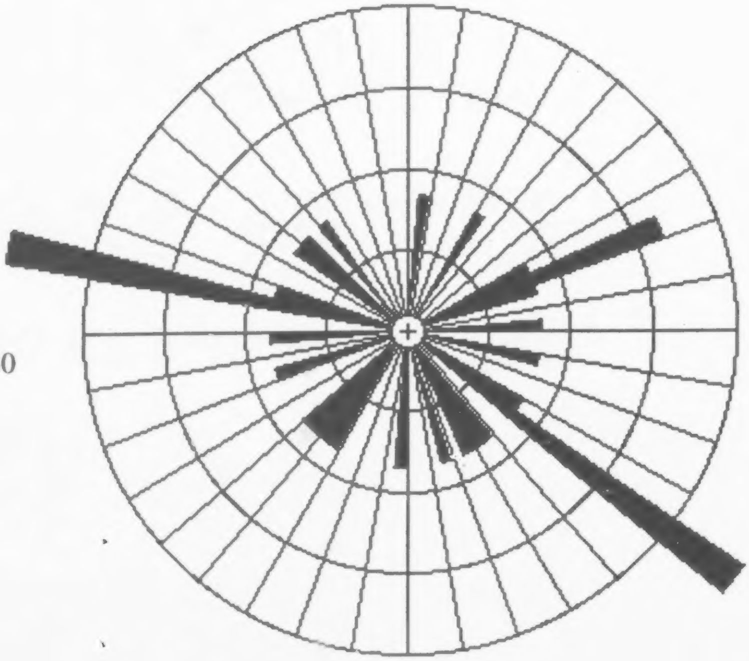
LC1-8
Unit - 2
Trend - 242.8
Plunge - 0.4
S_1 - 0.53
S_2 - 0.38
Clast Number - 25
Multimodal



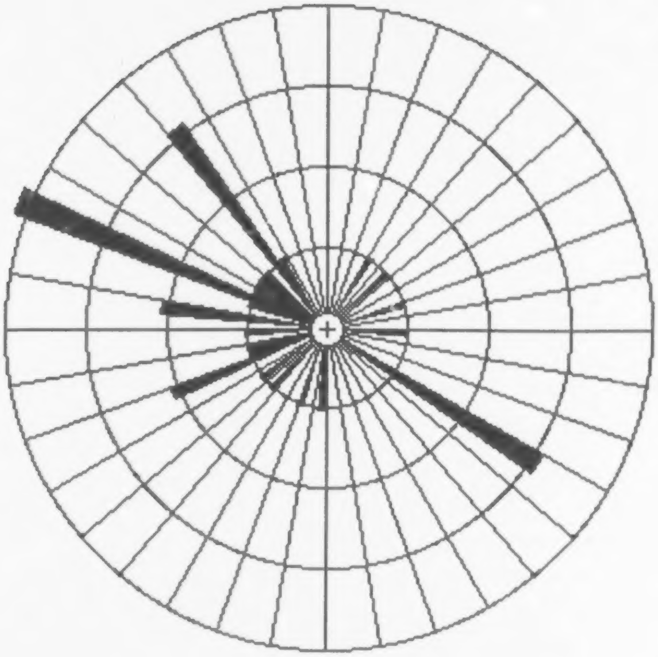
LC1-9
Unit - 2
Trend - 248.9
Plunge - 60.4
S_1 - 0.45
S_2 - 0.32
Clast Number - 25
Multimodal



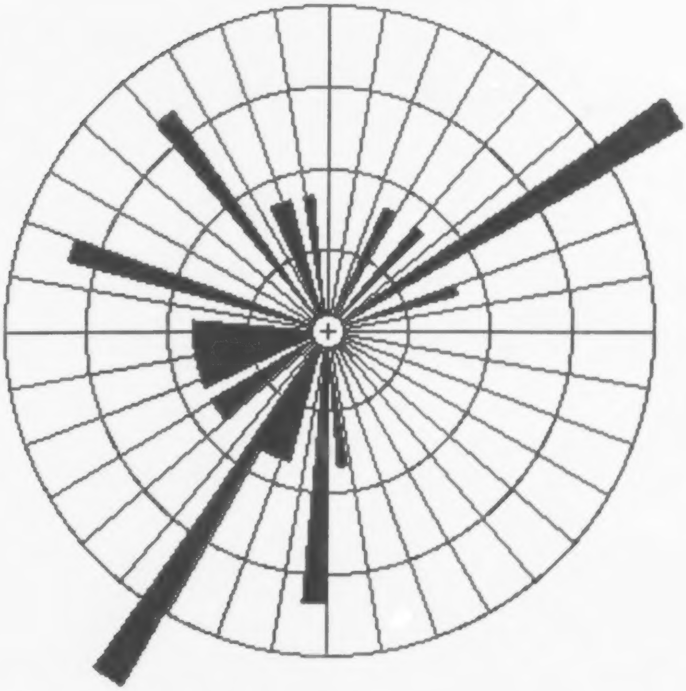
LC2-1
Unit - 1
Trend - 103.3
Plunge - 7.7
S_1 - 0.48
S_2 - 0.32
Clast Number - 30
Multimodal



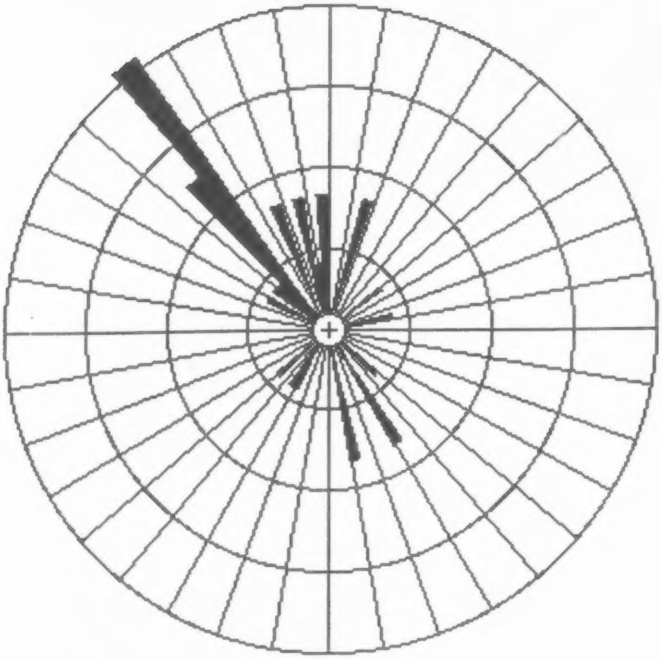
LC2-2
Unit - 1
Trend - 284.5
Plunge - 22.0
S_1 - 0.63
S_2 - 0.23
Clast Number - 31
Multimodal



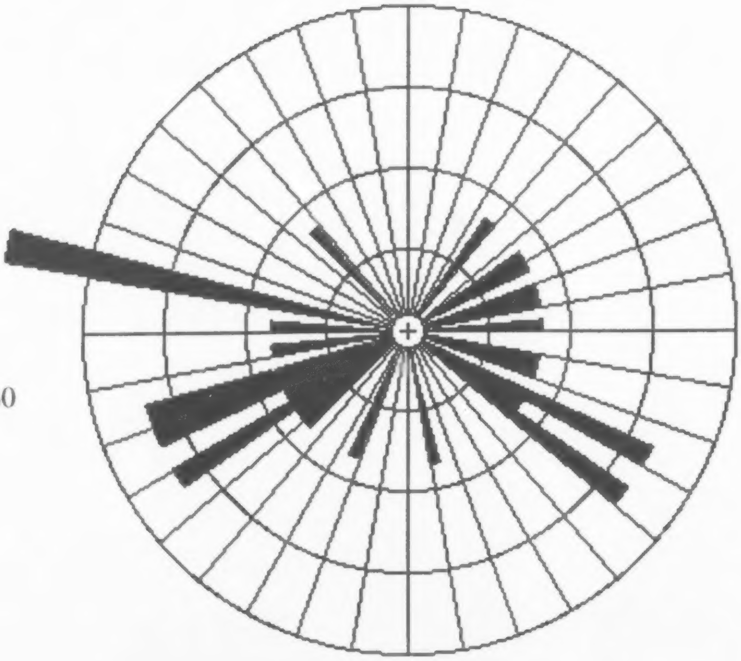
LC2-3
Unit - 1
Trend - 226.1
Plunge - 7.9
S_1 - 0.59
S_2 - 0.33
Clast Number - 30
Multimodal



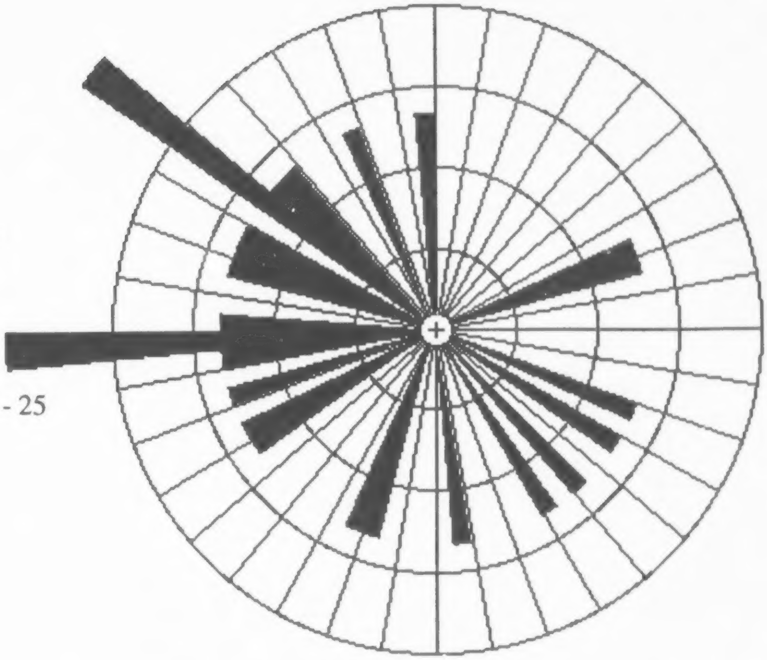
LC2-4
Unit - 1
Trend - 334.0
Plunge - 13.1
S_1 - 0.66
S_2 - 0.21
Clast Number - 30
Multimodal



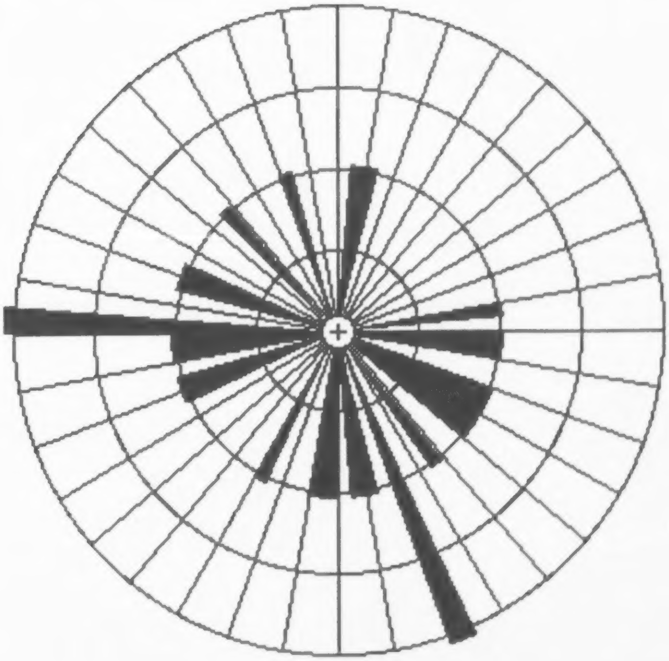
LC2-5
Unit - 1
Trend - 82.5
Plunge - 3.2
S_1 - 0.59
S_2 - 0.28
Clast Number - 30
Multimodal



LC2-6
Unit - 1
Trend - 288.3
Plunge - 9.4
S_1 - 0.58
S_2 - 0.29
Clast Number - 25
Multimodal

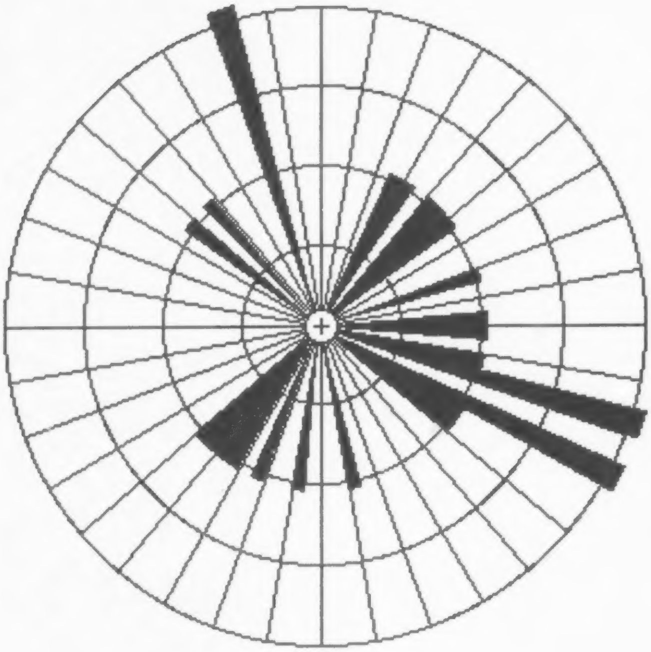


LC2-7
Unit - 2
Trend - 273.8
Plunge - 5.6
S_1 - 0.42
S_2 - 0.39
Clast Number - 27
Multimodal

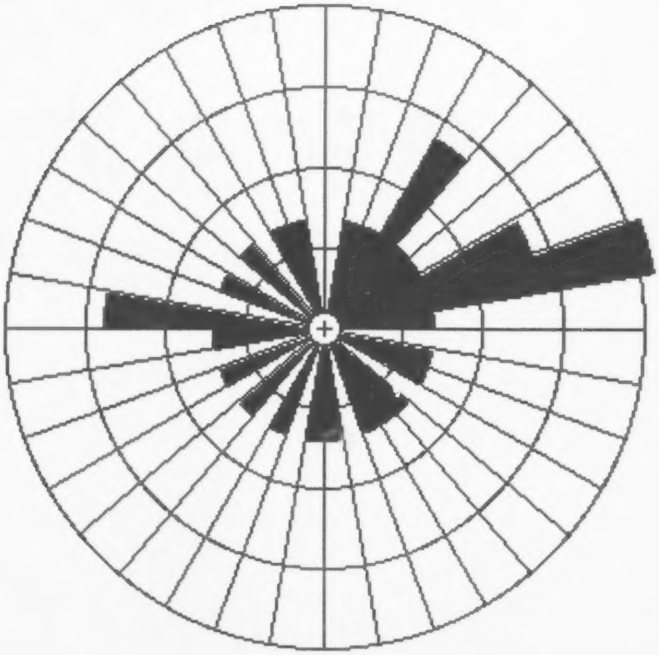


Rose Diagrams for Chapter 3

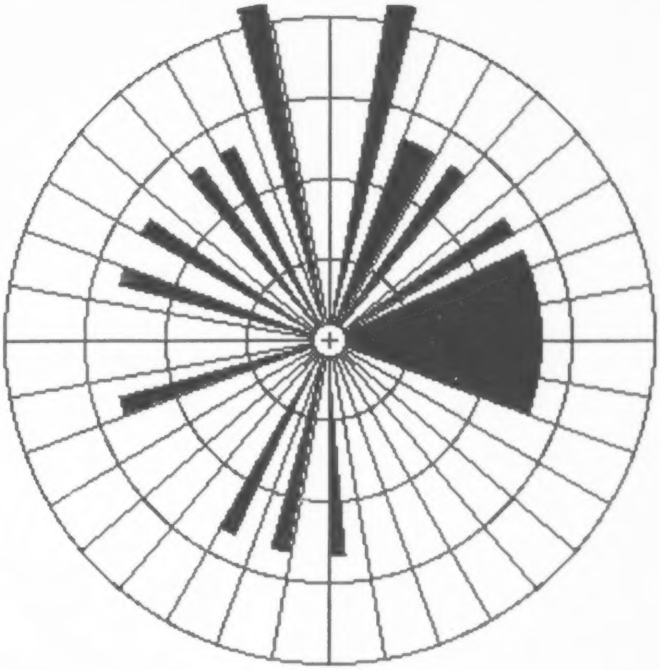
CS1-89
Unit - 1
Trend - 76.3
Plunge - 9.2
S_1 - 0.48
S_2 - 0.41
Clast Number - 26
Multimodal



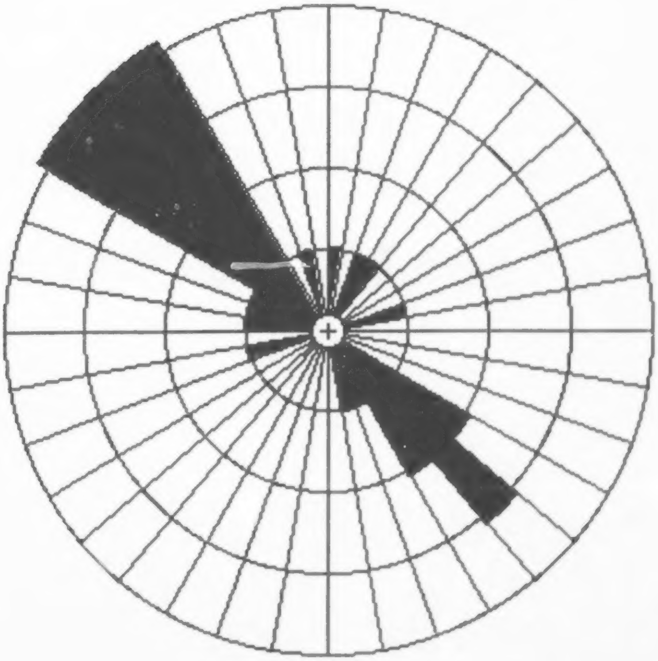
CS2-89
Unit - 1
Trend - 63.6
Plunge - 6.5
S_1 - 0.46
S_2 - 0.35
Clast Number - 26
Multimodal



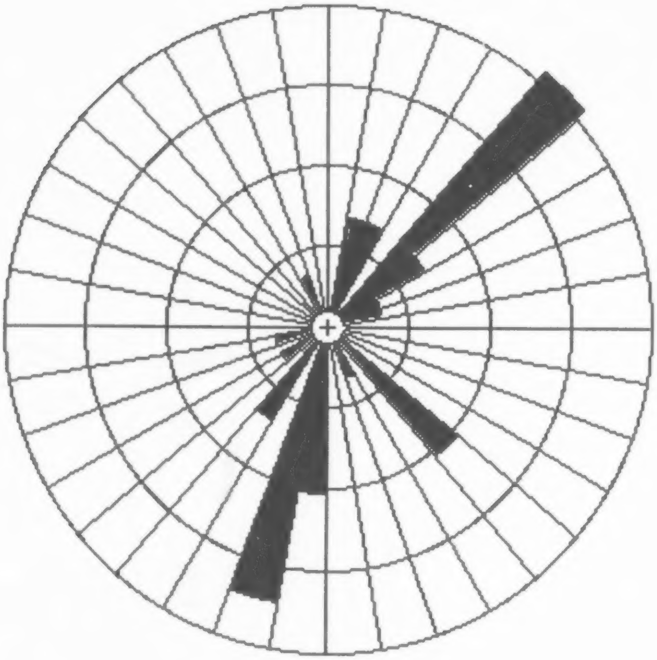
CS3-89
Unit - 1
Trend - 53.7
Plunge - 18.2
S$_1$ - 0.51
S$_2$ - 0.38
Clast Number - 25
Multimodal



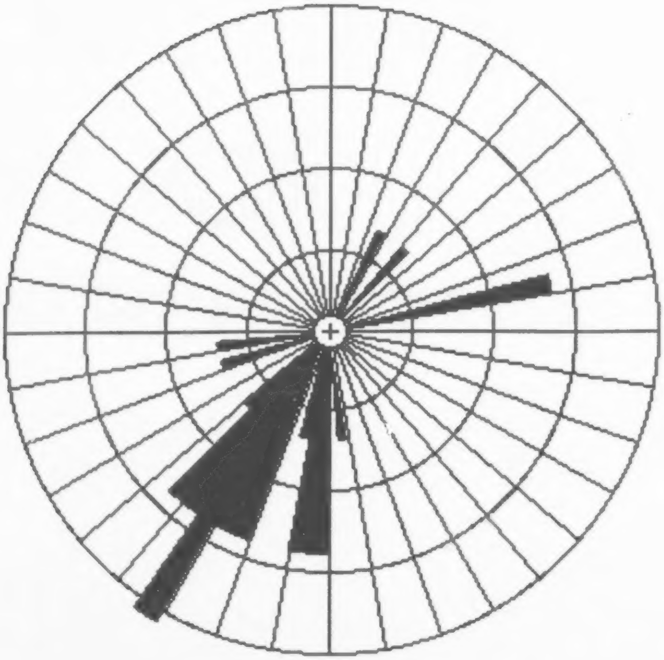
CS1-90
Unit - 2
Trend - 317.3
Plunge - 6.4
S$_1$ - 0.64
S$_2$ - 0.27
Clast Number - 30
Multimodal



CS2-90
Unit - 2
Trend - 298.9
Plunge - 1.0
S_1 - 0.67
S_2 - 0.23
Clast Number - 31
Multimodal



GIL1-89
Unit - 2
Trend - 211.2
Plunge - 2.7
S_1 - 0.83
S_2 - 0.14
Clast Number - 24
Multimodal



CL1W

Unit - 1

Trend - 339.4

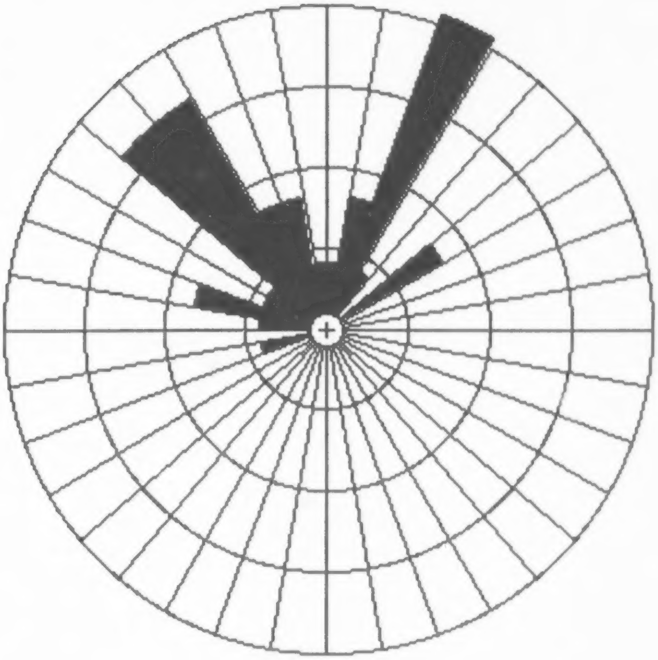
Plunge - 12.4

S_1 - 0.59

S_2 - 0.39

Clast Number - 22

Multimodal



CL2W

Unit - 1

Trend - 340.4

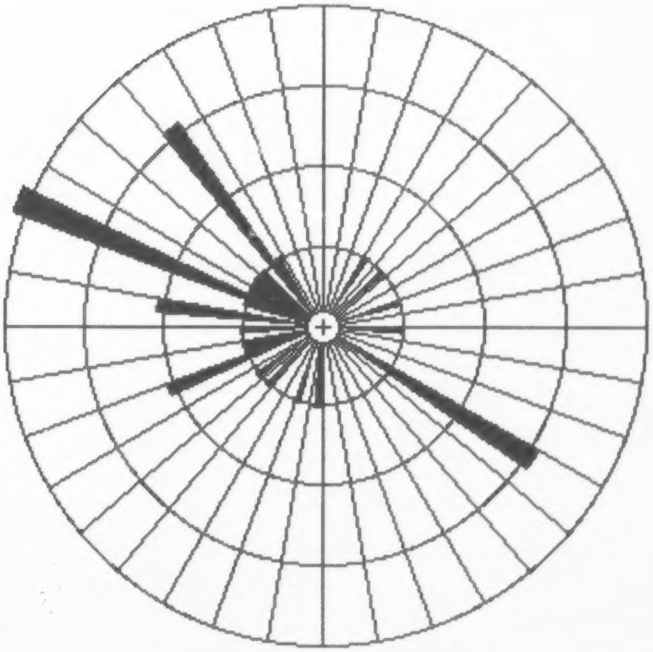
Plunge - 14.1

S_1 - 0.43

S_2 - 0.34

Clast Number - 30

Multimodal



CL3W

Unit - 1

Trend - 346

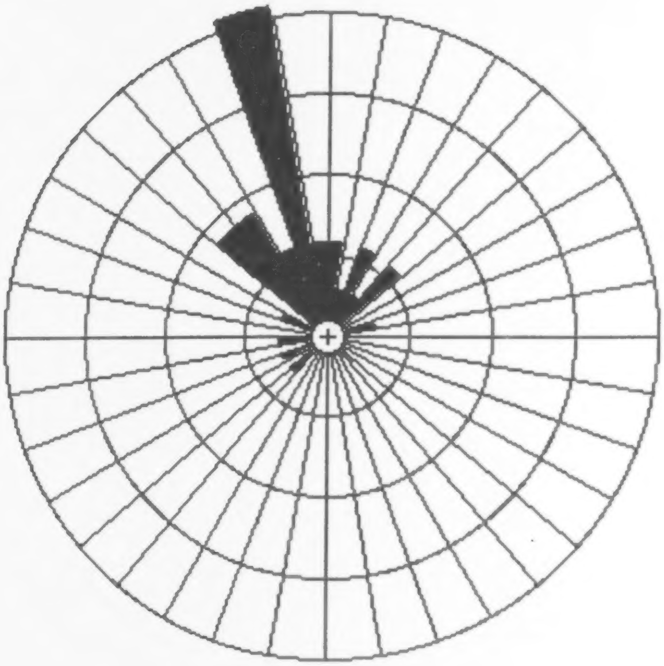
Plunge - 15.9

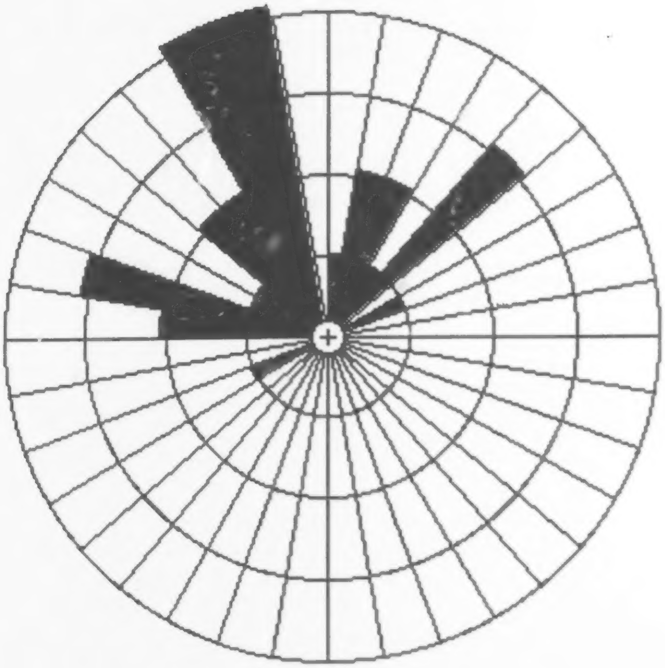
S_1 - 0.68

S_2 - 0.26

Clast Number - 30

Multimodal





CL6

Unit - 2

Trend - 337.6

Plunge - 11.3

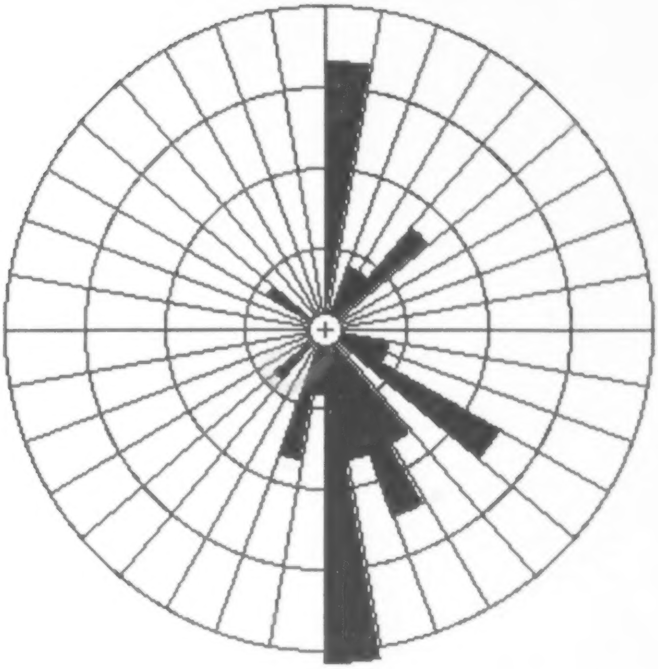
S_1 - 0.60

S_2 - 0.37

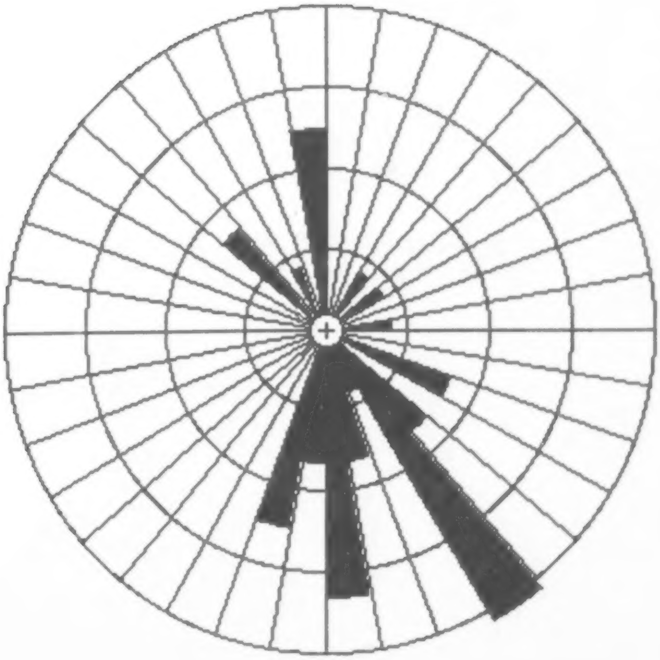
Clast Number - 30

Multimodal

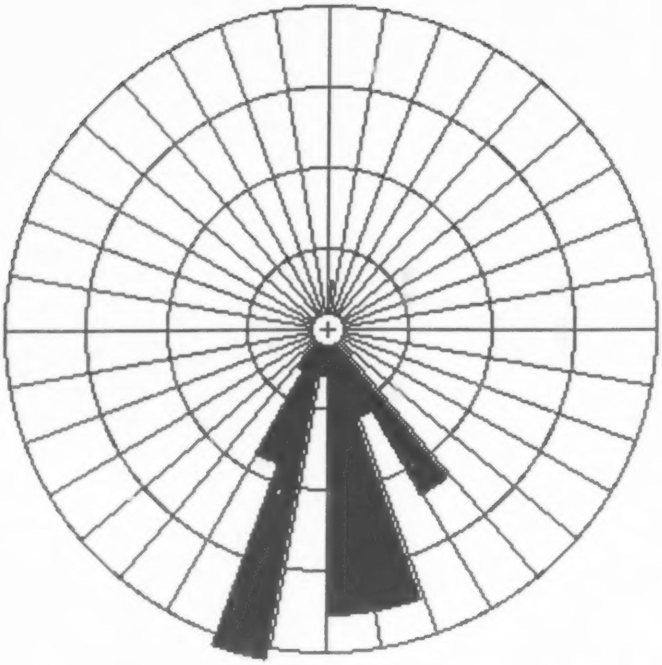
CL2E
Unit - 2
Trend - 165.8
Plunge - 9.4
S_1 - 0.48
S_2 - 0.43
Clast Number - 30
Multimodal



CL4E
Unit - 2
Trend - 160.9
Plunge - 10.0
S_1 - 0.52
S_2 - 0.38
Clast Number - 30
Multimodal



CL6W
Unit - 2
Trend - 177.1
Plunge - 13.3
S_1 - 0.48
S_2 - 0.30
Clast Number - 31
Multimodal



Appendix 3: Textural Data.

The textural data presented in the following tables provides the basic sand/silt/clay percentages of the bulk samples obtained from the sections described in Chapters 2 and 3.

Table A3.1. Sand/Silt/Clay percentages for Battle Creek 1.

Section	Sample number	Unit	Depth (m)	Sand	Silt	Clay	Median Phi
Battle Creek							
BC1	BC1-1	1	35	20.3	33.5	46.2	7.4
BC1	BC1-2	1	30	18.3	35.1	46.6	7.5
BC1	BC1-3	1	25	22.4	34.3	43.3	7.1
BC1	BC1-4	1	20	28.0	34.1	37.9	6.1
BC1	BC1-5	1	15	25.1	35.4	39.5	6.4
BC1	BC1-6	1	12	21.4	34.5	44.1	7.1
BC1	BC1-7	1	10	27.1	33.7	39.2	6.3
			average	23.2	34.4	42.4	
BC1	BC1-8	2	5	16.9	37.0	46.1	7.4
BC1	BC1-9	2	3	27.9	35.5	36.6	6.3
			average	22.4	36.3	41.3	

Table A3.2. Sand/Silt/Clay percentages for Lyons Creek 1.

Section	Sample number	Unit	Depth (m)	Sand	Silt	Clay	Median Phi
Lyons Creek							
LC1	LC1-1	1	18	8.3	52.7	38.9	6.3
LC1	LC1-2	1	16	22.4	35.4	42.2	6.0
LC1	LC1-3	1	14	28.2	32.6	39.2	6.3
LC1	LC1-4	1	12	28.3	32.3	39.4	6.3
LC1	LC1-5	1	10	29.8	32.7	37.5	6.1
LC1	LC1-6	1	8	24.1	34.3	41.6	6.7
LC1	LC1-7	1	6	24.1	34.8	41.1	6.6
			average	23.6	36.4	40.0	
LC1	LC1-8	2	4.5	24.1	34.2	41.7	6.7
LC1	LC1-9	2	2.5	23.1	33.8	43.1	7.0
LC1	LC1-10	2	1.5	25.6	34.3	40.1	6.5
			average	24.3	34.1	41.6	

Table A3.3. Sand/Silt/Clay percentages for samples from Lyons Creek 2.

Section	Sample number	Unit	Depth (m)	Sand	Silt	Clay	Median Phi
Lyons Creek							
LC2	LC2-1	1	22.5	19.6	31.7	48.7	7.8
LC2	LC2-2	1	18	21.2	32.7	46.1	7.4
LC2	LC2-3	1	16.5	23.4	31.8	44.8	7.2
LC2	LC2-4	1	14.5	25.2	33.4	41.4	6.7
LC2	LC2-5	1	12.5	27.3	34.3	38.4	6.3
LC2	LC2-6	1	11	26.7	33.4	39.9	6.5
LC2	LC2-7	1	9.5	26.6	34.3	39.1	6.4
LC2	LC2-8	1	7.2	27.4	33.5	39.1	6.4
LC2	LC2-9	1	6.0	27.0	34.0	39.0	6.3
LC2	LC2-10	1	5.0	27.4	33.7	38.9	6.4
LC2	LC2-11	1	3.5	26.3	34.6	39.1	6.4
			average	27.1	33.4	41.3	
LC2	LC2-12	2	3.0	27.4	29.6	39.4	6.4
LC2	LC2-13	2	2.2	27.6	31.0	41.4	6.6
LC2	LC2-14	2	1.3	27.3	34.1	38.6	6.2
			average	27.4	31.6	39.8	

Table A3.4. Sand/Silt/Clay percentages from the Canal Section.

Section	Sample number	Unit	Depth (m)	Sand	Silt	Clay	Median Phi
CANAL	CA-1	2	8.5	33.1	30.1	36.9	5.2
CANAL	CA-2	2	8.3	32.8	35.4	31.8	5.4
CANAL	CA-3	2	8.0	28.6	32.6	33.1	5.8
CANAL	CA-4	2	7.8	30.6	32.3	33.0	5.6
				31.3	32.6	33.7	
CANAL	CA-5	3	6.2	31.4	47.4	21.2	4.8
CANAL	CA-6	3	5.1	31.2	37.7	31.1	5.4
CANAL	CA-7	3	4.2	31.8	59.3	27.5	5.2
CANAL	CA-8	3	3.3	30.2	43.7	26.1	5.2
			average	31.2	47.0	26.5	
CANAL	CA-9(clay)	3	2.5	6.7	12.1	81.2	0
CANAL	CA-10 (silt diamicton)	3	2.0	10.2	45.3	45.5	7.4
CANAL	CA-11 (silt diamicton)	3	1.8	11.6	41.2	47.2	7.6
			average	10.9	43.3	46.4	

Table A3.5. Sand/Silt/Clay percentages from the Cypress Lake section.

Section	Sample number	Unit	Depth (m)	Sand	Silt	Clay	Median Phi
Cypress Lake							
CL	CL-1	1	5.5	42.9	27.4	29.7	4.7
CL	CL-2	1	5.1	25.4	34.7	39.9	6.6
CL	CL-3	1	4.7	25.2	32.6	40.4	6.5
CL	CL-4	1	5.2	20.4	50.7	28.9	6.1
CL	CL-5	1	4.3	31.0	32.7	37.4	6.0
			average	29.0	35.6	35.3	
CL	CL-6	2	3.8	37.1	33.6	29.3	5.8
CL	CL-7	2	3.5	20.4	28.2	51.4	7.8
CL	CL-8	2	2.5	24.2	36.3	39.5	6.4
CL	CL-9	2	2.1	42.1	31.1	26.8	4.8
			average	30.9	32.3	36.8	

Table A3.6. Gilchrist Ranch Section Textures.

Section	Sample Number.	Unit	Depth (m)	Sand	Silt	Clay	Median Phi
Gilchrist	GL-1	1 (clay band)	8	2.2	27.7	70.1	10
Gilchrist	GL-2	1 (clay band)	1	1.7	24.5	73.8	12
			average	1.9	26.1	71.9	
Gilchrist	GL-3	1 (silt bed)		5.6	66.3	27.7	7.8
Gilchrist	GL-4	1 (silt bed)		3.3	67.4	29.3	7.3
Gilchrist	GL-5	1		2.4	69.0	28.6	6.5
			average	3.8	67.6	28.5	
Gilchrist	GL-6	2	2.8	26.1	33.0	40.9	6.5
Gilchrist	GL-7	2	2.3	42.1	31.1	26.8	4.8
Gilchrist	GL-7	2	1.9	19.7	44.7	35.6	6.5
Gilchrist	GL-7	2	1.7	23.3	35.9	40.8	6.9
			average	27.8	36.2	36.0	

Appendix 4: Sites investigated in this study.

This appendix contains most of the sites and areas that were investigated and were formative in developing the ideas of the thesis. Figure A1 gives the locations of the sites. Township, range, and section designations for the site locations are also provided in the text. The sites are listed in broad west to east transects beginning in the northwest.

Site 1: Kame Site: West of 3, R28 T11 S36 1/4 Sect. 9.

The kame formed during the late stages of deglaciation of the area. It was probably located near the edge of the Etzikom ice.

Site 2: Delta area: West of 3, T11 R27 S31 1/4 Sect. 5.

The sand and gravel in this area are part of a delta which had prograded into a glacial lake located north of the Cypress Hills. The sand of the delta can be traced to the area south of the campground on Highway 1 northwest of Maple Creek.

Site 3: Lake sediments east of Maple Creek: West of 3, T12 R25 S16 1/4 Sect. 1.

This exposure along the southwest and northwest quadrants of quarter section 1 contains a thick sequence of silt and clay lake sediments. These were deposited in the lake north of Cypress Hills.

Site 4: Dunes near Piapot: West of 3, T12, R24, S15.

The sand dunes in this area are composed of sands deposited in the lake and exposed to wind reworking following drainage of the lake. Similar dunes are observed south of Highway 1 between Maple Creek and the Piapot dunes.

Site 5: Lake sediments northeast of Elkwater Lake, Alberta: West of 4, T8 R3 S27 1/4 Sect.. 9 and 10.

This exposure is located in a belt of hummocky terrain and contains overturned lake sediments. The exposure is immediately north of the Middle Creek margin. It can be debated whether it was formed during the advance of the Underdahl Ice as proposed by Vreeken (1986) or was deposited in a lake in front of the advancing Middle Creek ice (as proposed in Chapter 6).

Site 6: Paleosol on the West Block: West of 4, T8 R3 S15 1/4 Sect.. NW

The paleosol is located along the north side of the West Block above the terminus of the Laurentide ice. The area surrounding the paleosol has never been glaciated and no erratics are seen in this area of the West Block.

Site 7: Moraine 3 miles east of Reesor Lake: West of 4, T8, R1, S23, 1/4 Sect., 10 and 11.

The moraine extends south along a gap in the Cypress Hills. This moraine may mark the terminus of the Underdahl Advance in the area or it may reflect the terminus of the subsequent Middle Creek Advance. The uncertainty results from not knowing the location of the suture zone between the East and West lobes. It also reflects the view advanced in Chapter 6 that the East Lobe may not have advanced and retreated in the same sequence as the West Lobe.

Site 8: Moraine at Harris Lake: West of 3, T8 R29 S20 1/4 Sect.. 14 and 15,
Site 9: Moraine at Adams Lake: West of 3, T8 R29 S14 1/4 Sect.. 12 and 13,
Site 10: Moraine at Coulee Lake: West of 3, T8 R29 S12 1/4 Sect.. 6 and 11.

These lakes marks the terminus of the Laurentide ice north of the West Block of the Cypress Hills. They can be traced to the moraine at Reesor Lake and have the same problem associated with them.

Site 11: Gap Creek 1: West of 3, T9 R27 S19 1/4 Sect. SE of 9.
Site 12: Gap Creek 2: West of 3, T9 R27 S19 1/4 Sect. NE of 3.

Gap Creek 1 contains a thin unstratified clayey-silt till about 8 m thick which unconformably overlies shale bedrock. A 1 - 2 meters thick gravel covers much of the top of the section. The gravels are unusual as they are contorted and overturned. Gap Creek 2 contains clayey-silt till but there are numerous silt and fine sand layers which indicate that the sediments have been are pushed and overturned. This deformation may reflect readvance of the ice into the area. The readvance which caused the deformation (Middle Creek?, Altawan? or Pakowki? is uncertain). I believe they were overturned by the Pakowki Readvance but lacking dating control this is tentative.

Site 13: Lake sediments along Gap Creek: West of 3, T9 R27 S32 1/4 Sect.. Top half of 11.

Eight to ten meters of lake sediments are exposed in this section. The sediments consist of thick sequences of graded silts (30 to 140 cm thick silt beds composed of stacked 1 mm to 2 cm thick normally graded silt laminae and beds) interbedded with clay beds 5 to 10 cm thick. At the base of the sections there are clayey-diamicton layers containing a few erratic stones. These may be subaquatic debris flows. Similar exposures of lake sediments can be observed along Gap Creek to the east of this exposure. The lake sediments were most likely deposited in the same lake in which the delta to the north was emptying. It would therefore have formed during the retreat from the Pakowki Margin.

Site 14: Gravel pit in Cypress Hills Park, Saskatchewan: West of 3, T8 R27 S14 1/4 sect. southwest quadrant of No. 12.

This gravel pit is located near the eastern boundary of the Gap Lobe. In the area only a few scattered erratics are observed at the surface. These erratics are almost always confined to shallow channels which probably transported meltwaters to the south away from the ice to the north. This pit is interesting as there are periglacial involutions in it. These involutions are truncated by an upper 50 to 80 cm thick gravel unit. This gravel unit contains clasts imbricated by fluvial flow and is very sparse erratics. Glacial meltwaters therefore had reworked the upper part of the section.

Site 15: Push Moraines along Highway 21 south of Cypress Hills Park turnoff: W of 3, T8, R26, S 12, 1/4 sect., 3, 4, 5, and 6.

This is an area of hummocky gravels in the low between the East and West Blocks. These hummocky gravels mark the edge of the Late Wisconsinan Underdahl Advance. The gravels are derived from the pre-glacial Cypress Hills Formation. Scattered surface erratics are present but no erratics were found in roadcuts or in a gravel pit in this zone. It may be that the large volume of pre-glacial gravels and boulders has masked the erratics that are there. It is also possible that the thin ice was very clean and did not contain anything but locally derived materials. Why the ice did not extend further south across the low between the East and West Blocks is unclear. It may be that the cliff-like north flanks of these two blocks inhibited movement of the ice here. It may also relate to a general drop in maximum elevation reached by the ice as it is traced to the east (1420 m at Elkwater, Alberta, 1110 m here, and between 870 and 885 m elevation south of Rockglen, Sask. There the terrain was never glaciated. This drop in ice margin elevation to the east may be related to slowing of the ice as it ascended the Missouri Coteau.

Site 16: Gravel ridge west of Sucker Creek: West of 3, T7 R26 S10 1/4 sect, northeast corner of 9 southeast corner of 16.

This ridge on air photographs appears to be some form of moraine. Investigation on the ground reveals it to be a gravel ridge. No erratics were observed on the surface or in an excavation into the ridge. Examination of a bulk sample of coarse sand from the ridge revealed feldspar and granite grains. There is therefore are erratic component present but not in the gravel and boulder fractions. This again may reflect dilution of those fractions by incorporation of large quantities of pre-glacial Cypress Hills gravels.

Site 17: Shallow channels on Highway 21 south of the Cypress Hills Park turnoff: West of 3, T7, R26, S25, 1/4 sect. 9, 10, 11, 12.

These small, shallow channels (1-2 m deep 5 - 20 m across) were cut by meltwaters draining southwards from the northern ice mass. They are infilled by sediments deposited from the northern ice mass.

Site 18: Canal Section: West of 3, T7 R25 S19 1/4 sect, southeast quadrant of 8 to the southeast quadrant of 16.

This section extends about 1.25 km along an irrigation canal on the west flank of the East Block. The section contains unstratified clayey stoney diamicton (a subglacial meltout till or a thick subaquatic debris flow) overlain by a stratified unit of subaquatic glacigenic debris flows interbedded with clay and silt beds deposited by turbid underflows and suspension. This sediment occurs along the northern margin of the East Lobe where it ascended onto the East Block.

Site 19: Gravel ridge east of the Canal Section: West of 3, T7, R25, S23, 1/4 sect, northeast quadrant of 4, northwest quadrant of 3.

This ridge is similar to the on near Sucker Creek. On air photographs it appears to be an esker or some form of thin recessional moraine. Investigation on the ground reveals it to be a gravel ridge. No gravel or boulder-sized erratics were observed on the ridge itself. But examination of a bulk sample of coarse sand from the ridge again revealed feldspar and granite grains. Another case of dilution through the incorporation of large quantities of pre-glacial Cypress Hills gravels. This ridge was used to define the edge of the East Lobe in the area.

Site 20: Moraine in the Fairwell Creek area: West of 3, T7, R24, S1, 1/4 sect, 2 and 4; T7, R24, S2, 1/4 sect, 16; T7, R24, S11, 1/4 sect. 1.

This moraine consists of hummocky terrain east of Fairwell Creek at about 1110 to 1120 m elevation along the road which connects the Canal Section area to Ravenscrag. It marks the edge of the East Lobe where it ascended a steep slope.

Site 21: Ravenscrag Pit: West of 3, T7, R23, S7, 1/4 sect, northeast quadrant of 8 and the southeast quadrant of 9.

This pit is excavated into pre-glacial gravels of the Cypress Hills Formation and lies at 1145 m elevation. There is no glacigenic diamicton on the surface surrounding the pit but scattered granite and carbonate erratics are present. In the gravels of the pit, numerous periglacial involutions and ice wedge casts can be seen. In aerial photographs of this area, a channel system (Blacktail Creek) can be observed on the flat surrounding the pit. The scattered erratics may have been ice-rafted across glacial lake Blacktail.

Site 22: Dollard Drumlin Field: West of 3, T8, R19, S30, 1/4 sect, northeast quadrant of 4; T8, R19, S6, 1/4 sect, north half of 8; T7, R20, S26, 1/4 sect., southeast quadrant of 2 and southwest quadrant of 1; T7, R20, S20, 1/4 sect, 7; T7, R20, S19, 1/4 sect., southeast quadrant of 11;T7, R20, S21, S2, 1/4 sect., west half of 15; T7, R21, S11, 1/4 sect., east half of 2.

The above are examples of drumlins found in the Dollard Drumlin Field. The drumlins often consist of bedrock cores with a capping of sand and gravel. Shaw (1993, pers. comm.) has stated his belief that the drumlins were formed by sub-glacial meltwater erosion. The waters incising the drumlins were part of the mega-flood described in Rains et. al. (1994). An alternate proposition, advanced in Chapter 4, is that the drumlins were formed by sub-glacial processes reflecting active ice. The shallow lakes up-ice of many of the drumlins may have been the source of the bedrock in the drumlin cores. This would be similar to the hill - hole pairs described by Bluemle (1970) and Clayton and Moran (1974). The drumlins were then draped by sand and gravel deposited from the subglacial drainage of the large lake impounded north of the Cypress Hills.

Site 23: Manyberries Ash: West of 4, T5, R6, S13, 1/4 sect, northeast quadrant of 12, northwest quadrant of 11.

The Manyberries Ash was deposited in sediment covering ice derived from the Pakowki Advance. The Pakowki Advance therefore predates this ash layer. It is one of the Glacier Peak Ashes and is equivalent to either the Manyberries event (12,750 ±350 BP) or the Chiwawa event (11,200± 100 BP) (Vreeken 1986).

Site 24: Middle Creek Eskers: West of 3, T5, R30, S10, 1/4 sect, E half of 10 and E side of 15, T5, R30, S15, 1/4 sect., East half of 1 and SE of 8; T5, R30, S13, 1/4 sect.; SW to NE of 5, NW 11, SE 14.

These eskers enter the Middle Creek channel at grade through smaller channels. The eskers indicate the intimate relationship between the Middle Creek channel and the hummocky terrain to the south.

Site 25: Middle Creek slip-off slopes: West of 3, T5, R30, S25, 1/4 sect, 15; T5, R29, S29, 1/4 sect.; 2-3.

These slip-off slopes indicate that the Middle Creek channel is a normal fluvial channel and not a channel incised by sub-glacial mega-flood drainage as hypothesized by Shaw (1993, 1994, pers. comm.). The character of the channel difffers significantly from that of the Frenchman channel to the east which was incised by catastrophic flows and which does not contain similar slip-off slopes.

Site 25: Merryflat moraine: West of 3, T6, R29, S11; and S14, 1/4 sect. 1, 2, 3, and 4.

This moraine extends across Merryflat. It was deposited from the West Lobe during the Underdahl Advance at the Late Wisconsinan maximum. It marks the limit of the Late Wisconsinan ice in the area.

Site 26: Scroll bars on the interfluve between Adams and Battle creeks: West of 3, T6, R29, S34.

These diffuse scroll bars mark a pre-existing channel system on the flanks of the West Block. I believe they were formed during drainage of glacial lake Graburn which had covered the south slope of the West Block at the Late Wisconsinan maximum.

Site 27: Gravel pit along the Battle Creek: West of 3, T5, R28, S20, !/4 sect., 14 and 15.

This gravel pit is located along the base of the channel near the start of the Frenchman channel. The pit contains sand and gravel. Gravel clasts are imbricated and indicate flow from the west. The presence of "normal" fluvial gravels in this location indicates that flows along the Middle Creek channel were not catastrophic outbursts but ordinary fluvial flows. These flows post-dated formation of the Frenchman channel.

Site 28: Gravel pit north of the Battle Creek and east of Merryflat: West of 3, T5, R28, S29, 1/4 sect., 10.

This gravel pit contains glaciofluvial sand and gravel. It is situated on the flanks of the Center Block and contains well-developed foreset beds that are 10 to 20 m in length. These gravels may have been deposited in a delta and may be related to the early stages of formation of glacial lake Cypress. They clearly mark an earlier stage of meltwater flow above the level of the Frenchman channel and may have been the location of the initial outburst flow. Rapid erosion of the confining ice to the south would have shifted the channel to the location of the Frenchman channel.

Site 29: Hummocks in the Frenchman channel: West of 3, T5, R28, S34, 1/4 sect., 2 and east half of 3.

This small set of hummocks in the Frenchman channel indicate that the Middle Creek advance had reached this area. This and sites 26, 27, and 30 indicate that the Middle Creek advance had affected the initiation area of the Frenchman channel. It also indicates that the Middle Creek channel post-dates the Frenchman channel.

Site 30: Gravel pit west of Cypress Lake: West of 3, T6, R27, S6, 1/4 sect., 14.

This gravel pit contains glaciofluvial sand and gravel. It is situated in the Frenchman channel south and west of Cypress Lake. These gravels were derived from outflows from the Middle Creek advance. They eroded previous deposits and complicated the glacial history of the area.

Site 31: Cypress Lake sections: West of 3, T6, R27, S12, 1/4 sect., northeast quadrant of 6 and northwest quadrant of 7.

These sections are described in Chapter 3. The sections consist of a highly variable lower unit (perhaps a sub-glacial meltout till that has been modified or a glacigenic debris flow) overlain by a stratified unit of thin diamicton beds intercalated with sand and silt laminae. The second unit is a glaciolacustrine diamicton unit deposited in glacial lake Cypress.

Site 32: Gilchrist Ranch section: West of 3, T6, R26, S36, 1/4 sect., southwest quadrant of 4.

This section described in Chapter 3, contains glaciolacustrine silts and clays unconformably deposited over Cypress Hills Formation gravels. The gravels contain periglacial features. Over the silts and clays is a unstratified clayey-stoney diamicton. A transition zone separates the two units. Both units were deposited in a glacial lake formed when ice of the East Lobe advanced northwards onto the East and Center blocks.

Site 33: Streamlined bedforms in the Frenchman channel: West of 3, T6, R24, S27, 1/4 sect., top halves of 3 and 4.

These bedforms are observed in the Frenchman channel near Fairwell Creek and before its junction with Palisades Coulee. The forms are incised into bedrock. They maintain the stratigraphic sequence and so are unlikely to be slumps and appear to be streamlined. They may be similar to streamlined bedforms described by Kehew and Lord (1987). They may indicate that flow along this part of the channel was too short-lived to completely remove the bedrock.

Site 34: "Grooves" east of the Palisades Coulee: West of 3, T6, R27, S6, 1/4 sect., 14.

These grooves may indicate that during the early stages of drainage of Glacial lake Robsart, prior to incision of the Palisades Coulee, outflow from the lake covered a wide area and several small channels were cut. Following incision of the coulee these small channels were abandoned.

Site 35: Palisades Coulee eskers: West of 3, T5, R24, S16, 1/4 sect., northwest quadrant of 14 and southeast quadrant of 3; T5, R24, S15, 1/4 sect., northeast quadrant of 6 and southwest quadrant of 10.

These eskers drained the stagnant ice isolated around the Palisades Coulee following incision of the Frenchman channel. The meltwater from the eskers cut a channel into the base of the Palisades Coulee.

Site 36: Eastend gravel pits: West of 3, T6, R21, S30, 1/4 sect., 1 and 8; T6, R21, S29, 1/4 sect., 4 and south half of 5.

This is a set of gravel pits south of Eastend on the inside bend of the Frenchman channel where it turns to the south. The pits drop in elevation and contain boulders up to 1.5 X 2 X 1 m.

Site 37: Eastend kaolinite pit: West of 3, T6, R22, S13, 1/4 sect., southeast quadrant of 6.

This kaolinite pit contains a very good exposure of ice-thrust bedrock. The bedrock plunges to the northeast at a 25^0 angle. This indicates that ice of the East Lobe had extended to the southwest across the area of the Frenchman channel.

Site 38: Middle Creek section: West of 3, T4, R29, S20, 1/4 sect., southwest quadrant of 7.

This section consists of 6 to 8 m of clayey stoney diamicton unconformably overlying shale bedrock. The diamicton is dense and consolidated and maybe a basal till. This section was not described in detail in the thesis as it was a single section which could not be correlated to any nearby sections. It would be an interesting section for detailed analysis as it contains bedrock blocks and stringers which appear to be streaked out. Furthermore some partings and color breaks in the diamicton can be traced directly to 1- 4 mm thick bedrock stringers. This may mean deposition via lodgement or smudging while in the ice and preservation after meltout. I believe it is lodgement but more investigation is needed.

Site 39: Altawan scabland: West of 3, T3, R30, S16.

This is a broad zone in which erratics are concentrated on a shale bedrock surface. The scabland is bounded in some areas by what appear to be channel walls about 8 to 10 m high. Scattered throughout the scabland are erratic boulder concentrations and isolated glacigenic diamicton beds. Sand and gravel deposits were not observed. This scabland area occurs adjacent to the Altawan Advance margin and may have transported meltwaters impounded further to the northwest that had been impounded by the ice. This scabland zone connects up to the Medicine - Lodge channel in Alberta. The paucity of sand and gravel deposits indicate that the meltwater that flowed here was very erosive.

Site 40: Altawan Reservoir: West of 3, T2, R30, S35, 1/4 sect., east half of 12 and east half of 13; T3, R30, S1, 1/4 sect., west half of 2.

A number of sections are present on the southwest side of the Altawan Reservoir. These sections consist of diamicton layers (flow tills?) interbedded with sand and silt layers. Both the diamicton beds and the sorted sediments show a wide range of contortions which probably reflect the meltout of buried ice blocks. One diamicton section is capped by a 3 m thick sequence of climbing ripples in fine sands. These ripples were deposited by rapid deposition from suspension. The glaciofluvial meltwaters therefore must have been choked with sediment. The deposits were emplaced during retreat from the Altawan margin.

Site 41: Govenlock sections: West of 3, T3, R29, S23, 1/4 sect., southwest quadrant of 12; T3, R29, S23, 1/4 sect., northeast quadrant of 12; T3, R29, S23, 1/4 sect., northwest quadrant of 13.

These three sections consist of layers of silts and very fine sands interbedded with clayey stone-poor diamicton beds. In Govenlock 1 these beds have vertical or nearly vertical attitudes while in Govenlock 2 and 3 the beds though contorted are not as vertical. These sections are in the sediments that have partially infilled the early Middle Creek channel. This channel, visible in air photographs, transported meltwaters southwards probably from the battle Creek. Later, following retreat of the Middle Creek ice, a new channel was cut by the Middle Creek. The contortions and vertical beds indicate that the sediments were deposited over ice blocks. The sediments were probably deposited during retreat from the Middle Creek margin.

Site 42: Govenlock gravel pit: West of 3, T3, R29, S15, 1/4 sect., 16.

This gravel pit contains outwash sand and gravel deposited during retreat from the Middle Creek margin.

Site 43: Consul sections: West of 3, T4, R27, S12, 1/4 section, northwest quadrant of 12; T4, R27, S12, 1/4 section, northeast quadrant of 10.

These sections contain glaciolacustrine sediments (normally graded silt beds, clay beds and laminae and dropstones) that were deposited in glacial lake Consul. Glacial lake Consul formed during retreat from the Middle Creek margin.

Site 44: Ice thrust bedrock along Highway 46 between Divide and Clayton, Saskatchewan: West of 3, T2, R23, S26 - 30.

Along Highway 46 numerous roadcuts expose sandstone and shale bedrock that has been thrust by ice moving from the northeast. This indicates that the East Lobe had extended to the southwest at least to this area which includes the southeast flank of the Old Man On His Back Plateau.

Site 45: Staynor Hall Region: western edge of the confluence area between the East and West Lobes: Recticulate moraine: West of 3, T1, R23, South half of 6, section 7, northwest quarter of 8, sections, 17, 18, 19, west half of 20, southwest quarter of 30.

Moraine 2: West of 3, T1, R24, S22, 1/4 sect. 15, 16, T1, R24, S23, 1/4 sect., 13, 14, 15, 16, T1, R24, S25, 1/4 sect., 4, southwest quadrant of 5, T1, R24, S26, 1/4 sect. 1, T1, R24, S27, 1/4 sect. 2.

Moraine 3: West of 3, T1, R23, S35, 1/4 sect. 1, 2, 7, 8, east half of 6, T1, R24, S36, 1/4 sect., 2, 3, 4, 8, and southeast quadrant of 7, T1, R 23, S31, 1/4 sect., 12, and the southwest quadrant of 13.

These three moraines southwest of Staynor Hall probably mark the eastern limit of the West Lobe. The area contains little in the way of exposures. Glacigenic sediment is thin and in many areas shale bedrock is at or near the surface. The reticulate moriane is well-expressed in air photgraphs but on the ground it appears as an area of shallow poorly-developed hummocky terrain. This moraine lies on the west flank of the Boundary Plateau. Moraines 2 and 3 are arcuate in shape and have a braided appearance in air photographs. They may be end or recessional moraines formed at the edge of the West Lobe. A lack of funds prevented the drilling and backhoe excavation that was planned for this area. But it is the most likely location of the margin between the two ice lobes.

Site 46: Coriander section: West of 3, T3, R11, S31, 1/4 sect., 11.

This section consists of stratified glacigenic diamictons overlying shale bedrock. The section is a single isolated section that is 7 m high. The sediment was probably deposited by glacigenic debris flows. Thin section analysis of diamicton from the unit did not reveal any features that could conclusively identify the origin of the deposit.

Site 47: Battle Creek sections:

BC1: West of 3, T1, R26, S4, 1/4 sect., South half of 3 to SW quad. of 2,
BC2: T1, R26, S4, 1/4 sect., SW quadrant of 1,
BC3: T1, R26, S4, 1/4 sect., NE quadrant of 4 to SW quadrant of 6,

BC4: T1, R26, S21, 1/4 sect., SE quadrant of 2,
BC5: T1, R26, S33, 1/4 sect., SW quadrant of 13 to NE quadrant of 13.

These sections contain subglacial meltout till overlain by a subaerial glacigenic debris-flow assemblage (as described in Chapter 2). The sections contain the same units and were deposited during the Underdahl Advance. The terrain around the sections is flat to very gently undulating. In some areas scour channels from meltwater streams are evident. The top of BC4 is interesting in that the top of it has been eroded by the Battle Creek when it was flowing at the prairie level. The sand in the upper 2 m of the section are not deformed or faulted indicating that the ice from which Unit 1 (subglacial meltout till) was deposited has already melted. The Battle Creek has been linked in Chapter 6 to retreat from the Middle Creek margin. The time gap between the end of the Underdahl and the retreat of the Middle Creek was sufficiently long to allow all the ice to melt. As buried ice can persist for several thousands of years (see references in Chapters 2 and 6), the time interval between the two events was have been over several thousand years.

Site 48: Lyons Creek sections:
LC1: West of 3, T1, R25, S14, 1/4 sect., NW quadrant of 14,
LC2: T1, R25, S23, 1/4 sect., SW quadrant of 3,
LC3: T1, R25, S23, 1/4 sect., NE quadrant of 4,
LC4: T1, R25, S23, 1/4 sect., SE and NE quadrants of 11.

These sections contain the same subglacial meltout till and subaerial glacigenic debris-flow assemblage units as the Battle Creek sections (described in Chapter 2) and were deposited during the Underdahl Advance. The terrain around the sections is also flat to very gently undulating. In some areas scour channels from meltwater streams are evident. The Lyons Creek sections contain sediments deposited by the West Lobe. No useful sections were observed east of Lyons Creek.

Site 49: Hummocky terrain on the east flanks of the Boundary Plateau:
This is a broad belt centered around T1 R21 west of the 3rd meridian.

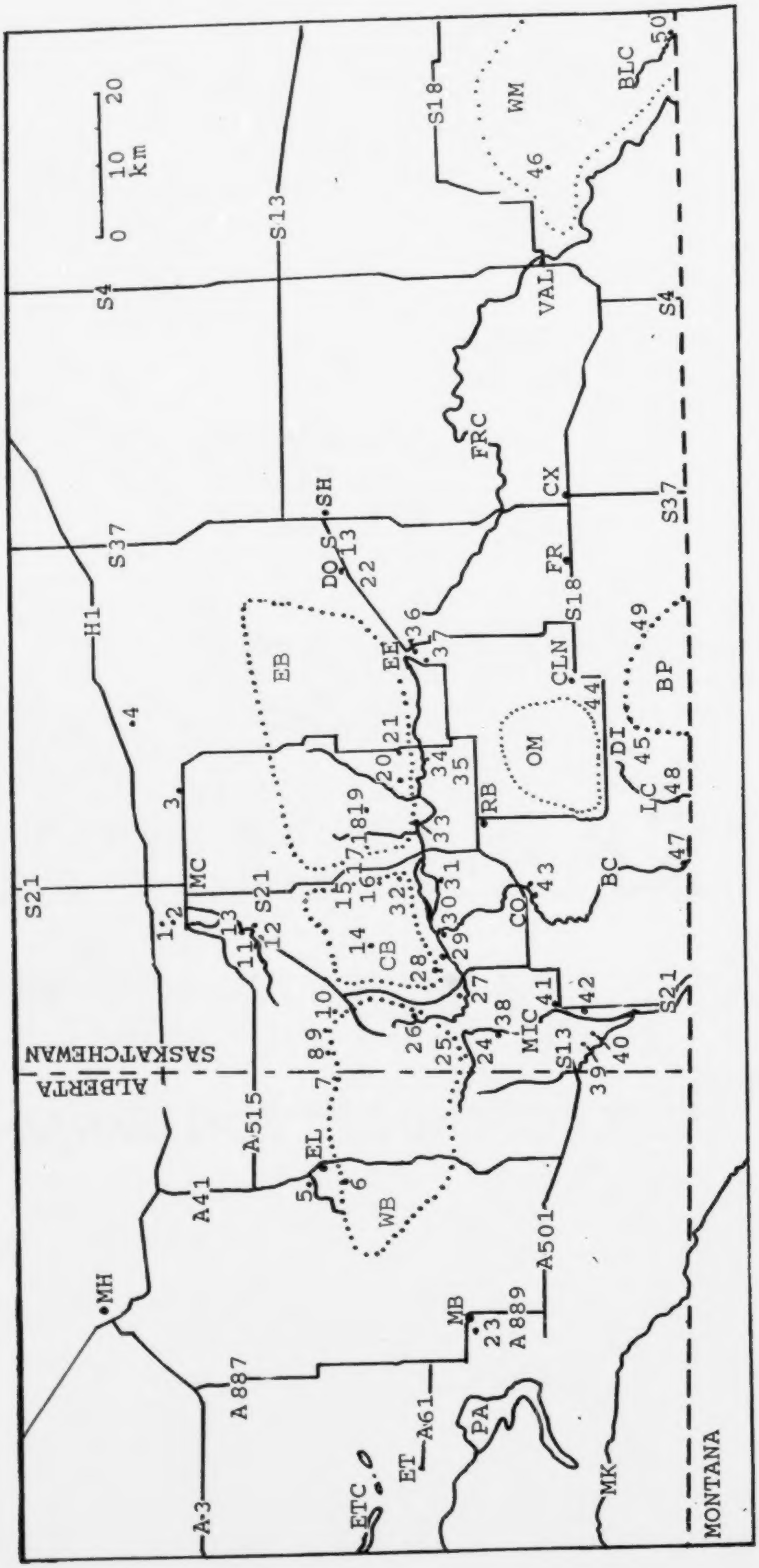
The hummocky terrain observed here has numerous ridges and hummocks that show that ice of the East Lobe flowing from the northeast, wrapped around the Boundary Plateau. This ice also extended onto the plateau and numerous channels are seen in the bedrock there. There is no indication that the West Lobe ascended onto the west side of the plateau. In fact as described for Site 45 (Staynor Hall) there is weak evidence indicating that the East Lobe had crossed over the plateau to very near its western margin. This is an area that may yield information on the pattern of earlier ice flow events but much of this information would be on the Montana part of the Boundary Plateau. This would be an intriguing project.

Site 50: Bluff Creek section: West of 3?, T1, R9, S10, 1/4 sect., 5.

The Bluff Creek section on the Wood Mountain map sheet (72G) contains a deep chocolate brown stoney diamicton overlying a gray brown stoney diamicton. This section has very little context and cannot be easily correlated to any of the poorly exposed sections in the area. It is interesting in that it contains the only exposure where a much older till may be exposed. This is a section that definitely needs more investigation.

Figure A4.1, Locations of sites investigated for this thesis.

BC- Battle Creek, BLC- Bluff Creek, BP- Boundary Plateau, CB- Center Block, CLN- Clayton, CO- Consul, CX- Climax, DI- Divide, DO- Dollard, EB- EAst Block, EE- Eastend, EL- Elkwater, ET- Etzikom, ETC- Etzikom Coulee, FRC- Frenchman Channel, LC- Lyons Creek, MB- Manyberries, MC- Maple Creek, MIC- Middle Creek, MH- Medicine Hat, MK- Milk River, OM- Old Man On His Back Plateau, PA- Pakowki Lake, RB- Robsart, SH- Shaunavon, VAL- Val Marie, WB- West Block, WM- Wood Mountain Upland. A3, A887, S13, S37 etc. are highways and secondary roads in Alberta and Saskatchewan. H1 is the Trans-Canada Highway 1.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References cited.

Bluemle, J.P., 1970, Anomalous hills and associated depressions in central North Dakota, Geological Society of America, Abstracts with Program, v. 2, p. 325-326.

Clayton, L., and Moran, S., 1974, A glacial-process form model, *in* Coates, D.R., (ed.), Glacial Geomorphology, Binghampton, New York, SUNY-Binghampton Publications in Geomorphology, p. 89-119.

Kehew, A.E., and Lord, M.L., 1987, Glacial-lake outbursts along the mid-continent margins of the Laurentide ice-sheet, *in* Mayer, L., and Nash, D., eds., Catastrophic Flooding, Allen and Unwin, Boston, p. 95-120.

Vreeken, W.J. 1986, Quaternary events in the Elkwater area of southeastern Alberta: Canadian Journal of Earth Sciences, v. 23, p. 2024-2038.